Glaciers and Climate Change
–
Spatio-temporal Analysis of Glacier Fluctuations in the European Alps after 1850

Dissertation
zur
Erlangung der naturwissenschaftlichen Doktorwürde
(Dr. sc. nat.)

vorgelegt der
Mathematisch-naturwissenschaftlichen Fakultät
der
Universität Zürich

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Zürich, 2006

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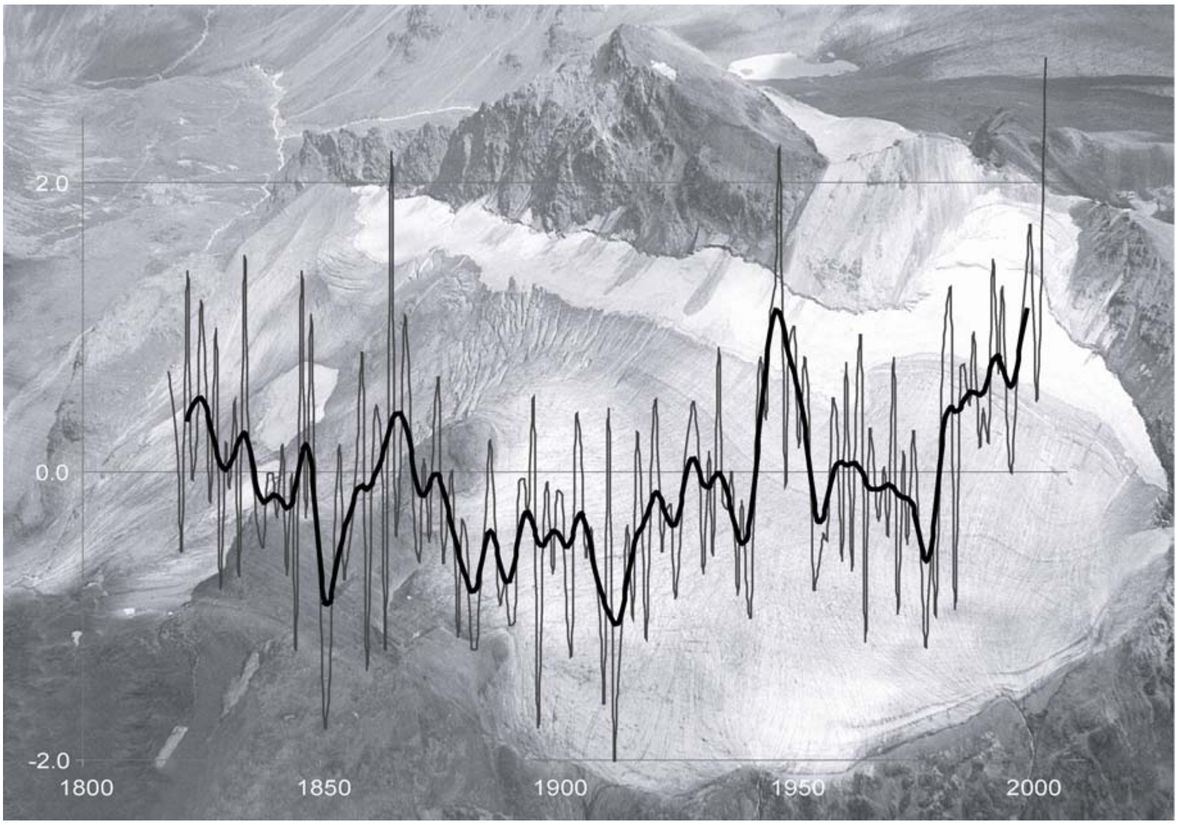

Michael Zemp

2006

Cover picture:
Calderas Glacier in the Upper Engadine, eastern Switzerland on August 9, 2003 (aerial photo kindly provided by C. Rothenbühler), overlaid with mean 6-month summer temperature anomalies in degree Celsius (single years and 10-years lowpass) to the 20th century mean of Alpine weather station above 1,500 m a.s.l. (provided by the Central Institute for Meteorology and Geodynamics in Vienna, Austria).

Summary

In densely populated high mountain regions such as the European Alps, glaciers are an inherent component of the Alpine culture, landscape and environment. They represent a unique resource of fresh water for agriculture and industry, an important economic component of tourism and hydro-power production, and a potential source of serious natural hazards. Due to their proximity to the melting point, glaciers are considered among the best natural indicators of global climate change. Mountain glaciers have become the leading icon in the current debate on climate change and on the uniqueness of present-day changes as compared to the variations which occurred during the Holocene period. Although numerous studies have investigated the relationship between glaciers and climate change, glacier-climate studies that focus on an entire mountain range and integrate in-situ measurements, remote sensing data and modelling approaches, as proposed by modern monitoring strategies, have for the most part been lacking.

It is to fill this gap in our understanding that this study undertakes to investigate glacier fluctuations in the European Alps after 1850. Glacier inventories, in-situ measurements and a numerical model (based on an empirical relationship between precipitation and temperature at the glacier steady-state equilibrium line altitude) are used in combination with a digital elevation model and GIS techniques to analyse the glacier fluctuations between 1850 and the end of the 21st century of the entire Alpine mountain range.

Overall area loss after 1850 is calculated using different methods as being about 35% until the 1970s, when the approximately 5,150 Alpine glaciers covered a total area of 2,909 km^2. This loss reached almost 50% by 2000. Rapidly shrinking glacier areas, spectacular tongue retreats and increasing mass losses are clear signs of the atmospheric warming observed in the Alps during the last 150 years and its acceleration over the past two decades, culminating in an ice loss of another 5–10% of the remaining ice volume during the extraordinary warm year of 2003. From the model experiment it is found that for Alpine glaciers, a change of ±1 °C in 6-month summer temperatures would be compensated by an annual precipitation increase/decrease of about 25%. A summer temperature rise of 3 °C would reduce the glacier cover of the reference period (1971–1990) by some 80%, or up to 10% of the glacier extent of 1850. In the event of a 5 °C summer temperature increase, the Alps would become almost completely ice-free. Annual precipitation changes of ±20% would modify such estimated percentages of remaining ice by a factor of less than two.

The presented study demonstrates how modern monitoring strategies can be applied for the investigation of glaciers of an entire mountain range, and that the probability of glaciers in the European Alps disappearing within the coming decades is far from slight.

The thesis consists of a collection of five papers; the studies reported on were carried out within the framework of the EU-funded ALP-IMP project and of the World Glacier Monitoring Service.

Zusammenfassung

Gletscher sind in dicht besiedelten Hochgebirgsregionen, wie den Europäischen Alpen, ein fester Bestandteil der alpinen Kultur, Landschaft und Umwelt. Sie sind eine bedeutende Süsswasserressource für die Landwirtschaft und die Industrie, eine wichtige wirtschaftliche Komponente für den Tourismus und die Produktion von Wasserkraftenergie und eine potentielle Ursache für Naturgefahren. Innerhalb der globalen Klimabeobachtungssysteme zählen Gletscher, durch ihre physikalische Nähe zum Schmelzpunkt, zu den besten natürlichen Klimaindikatoren. Gebirgsgletscher wurden zum Leitsymbol der gegenwärtigen Diskussion zum Klimawandel und zur Einzigartigkeit der aktuellen Veränderungen im Vergleich zur Holozänen Variabilität. Zahlreiche Studien haben die Beziehung zwischen Gletscher und Klimaänderung untersucht. Gleichwohl existieren nur wenige Gletscher-Klima Studien, die sich auf eine gesamten Gebirgskette konzentrieren und in einem integrativen Ansatz Feldmessungen, Fernerkundungsdaten und Computermodelle anwenden, wie es in internationalen Monitoring-Strategien vorgeschlagen wird.

Die vorliegende Dissertation untersucht daher Gletscheränderungen nach 1850 in den gesamten Alpen. Gletscherinventare, Feldmessungen und ein numerisches Model (basierend auf einer empirischen Beziehung zwischen Niederschlag und Temperatur auf der Höhe der Gleichgewichtslinie) werden verwendet, in Kombination mit einem digitalen Höhenmodel und GIS-Technologie, für die Analyse der Gletscheränderungen der Gesamtalpen zwischen 1850 und dem Ende des 21. Jahrhunderts.

Der gesamte Flächenverlust seit 1850, berechnet mit verschiedenen Methoden, beläuft sich auf etwa 35% bis in den 1970er Jahren, als ca. 5'150 Alpengletscher eine Fläche von 2'909 km^2 bedeckten, und auf fast 50% bis im Jahre 2000. Rapide schwindende Gletscherflächen, spektakuläre Rückzüge der Gletscherzungen und zunehmende Massenverluste sind klare Zeichen der atmosphärischen Erwärmung in den Alpen während den vergangenen 150 Jahren und deren Beschleunigung in den letzten zwei Jahrzehnten, die in einem Verlust von weiteren 5–10% des verbleibenden Eisvolumens im ausserordentlich warmen Jahr 2003 gipfelte. Aus dem Modellierungsexperiment wurde ersichtlich, dass es für die Kompensation einer Änderung der 6-Monats-Sommertemperatur um ±1 °C eine Zu-/Abnahme des jährlichen Niederschlages von etwa 25% benötigt. Ein Anstieg der Sommertemperatur um 3 °C würde die Alpine Gletscherbedeckung der Referenzperiode (1971–1990) um ungefähr 80% reduzieren, oder auf ca. 10% der Gletscherausdehnung um 1850. Im Falle eines Anstieges der Sommertemperatur um 5 °C würden die Alpen praktisch eisfrei werden. Jährliche Niederschlagsänderungen um ±20% modifizieren solche geschätzte Prozentwerte des verbleibenden Eises um weniger als ein Faktor Zwei.

Die vorliegende Studie demonstriert die mögliche Anwendung von modernen Monitoring-Strategien zur Untersuchung von Gletschern einer gesamten Gebirgskette und zeigt, dass die Möglichkeit des Verschwindens der Alpengletscher in den kommenden Jahrzehnten in Betracht gezogen werden muss.

Contents

Part A: Overview

Part B: Papers

The thesis is based on five papers, referred to in the text by roman numerals. Please note that they are ordered according to thematic considerations. A complete list of publications with contributions by the PhD candidate is given on pages 63–64.

I **Zemp, M.**, Paul, F., Hoelzle, M. & Haeberli, W. (in press): *Glacier fluctuations in the European Alps 1850–2000: an overview and spatio-temporal analysis of available data*. In: Orlove, B., Wiegandt, E. & Luckman, B. (eds.): The darkening peaks: Glacial retreat in scientific and social context. University of California Press.

II Hoelzle, M., Chinn, T., Stumm, D., Paul, F., **Zemp, M.** & Haeberli, W. (accepted): *The application of glacier inventory data for estimating past climate-change effects on mountain glaciers: a comparison between the European Alps and the Southern Alps of New Zealand*. Global and Planetary Change, special issue on climate change impacts on glaciers and permafrost.

III **Zemp, M.**, Frauenfelder, R., Haeberli, W. & Hoelzle, M. (2005): *Worldwide glacier mass balance measurements: general trends and first results of the extraordinary year 2003 in Central Europe*. Data of Glaciological Studies [Materialy glyatsiologicheskih issledovanii], 99: p. 3–12.

IV **Zemp, M.**, Hoelzle, M. & Haeberli, W. (accepted): *Distributed modelling of the regional climatic equilibrium line altitude of glaciers in the European Alps*. Global and Planetary Change, special issue on climate change impacts on glaciers and permafrost.

V **Zemp, M.**, Haeberli, W., Hoelzle, M. & Paul, F. (submitted): *Alpine glaciers to disappear within decades?* Geophysical Research Letter.

List of Figures

List of Abbreviations

AAR	Accumulation Area Ratio of a glacier
AAR_0	steady-state Accumulation Area Ratio of a glacier
AD	Anno Domini
ALP-IMP	EU research project, focusing on climate variability in the ALPs based on Instrumental data, Model simulations and Proxy data
cAA	climatic Accumulation Area of a glacier
CO_2	carbon dioxide
DEM	Digital Elevation Model
ELA	Equilibrium Line Altitude of a glacier
ELA_0	steady-state Equilibrium Line Altitude of a glacier
ESRI	Environmental Systems Research Institute, GIS vendor
EU	European Union
GCOS	Global Climate Observing System
GHOST	Global Hierarchical Observing Strategy
GIS	Geographical Information Systems / Science
GLIMS	Global Land Ice Measurements from Space project
GTN-G	Global Terrestrial Network for Glaciers
GTOS	Global Terrestrial Observing System
ICSI	International Commission on Snow and Ice
IPCC	Intergovernmental Panel on Climate Change
LGM	Last Glacial Maximum, about 21,000 y BP
LIA	Little Ice Age, period of major Holocene glacier re-advances, peaking around 1850
LP DAAC	Land Processes Distributed Active Archive Centre
PRUDENCE	EU research project, focusing on the prediction of regional scenarios and uncertainties for defining European climate change risks and effects
PSFG	Permanent Service on Fluctuations of Glaciers
$rcELA_0$	regional climatic steady-state Equilibrium Line Altitude of a glacier
TTS/WGI	Temporary Technical Secretary for the World Glacier Inventory
UNEP	United Nations Environment Programme
USGS EROS	United States Geological Survey centre for Earth Resources Observation and Science
WGMS	World Glacier Monitoring Service
y BP	years Before Present, conventional ^{14}C, uncalibrated
YD	Younger Dryas, about 11,000–10,200 y BP

Part A: Overview

1 Introduction

The present study was carried out within the framework of the EU-funded ALP-IMP project (http://www.zamg.ac.at/ALP-IMP/) and of the World Glacier Monitoring Service (WGMS; www.wgms.ch). ALP-IMP is an integrated research project that focuses on multi-centennial climate variability in the **ALP**s based on **I**nstrumental data, **M**odel simulations and **P**roxy data. The present thesis was part of Work Package 4, focusing on glaciers as climate proxy and leading symbol of climate change. The research represented a unique opportunity to apply modern glacier monitoring concepts to the European Alps where, by far, the most concentrated amount of information about glacier fluctuations over the last one and a half century is available.

1.1 Motivation

Glaciers, ice caps and continental ice sheets cover some 10% (or 16×10^6 km^2) of the earth's land surface at the present time and covered about three times this amount during the ice ages (Paterson 1994). This corresponds to about 3% (or about 36×10^6 km^3) of the total water on earth and 75–80% of the world's total freshwater resources (Reinwarth & Stäblein 1972). If all land ice melted away, the sea level would rise by about 81 m, with the Antarctic Ice Sheet contributing 73 m, the Greenland Ice Sheet about 7.5 m and all other glaciers and ice caps about 0.5 m to this rise (Meier & Bahr 1996, Zuo & Oerlemans 1997a, Oerlemans, 2001). Meier & Bahr (1996) estimated the total number and area of these glaciers and ice caps to be about 160,000 and 680,000 km^2, respectively. Recent estimates by Dyurgerov & Meier (2005) are somewhat larger, with 785,000 km^2. The 5,154 glaciers in the European Alps with a total area of 2,909 km^2 in the 1970s (IAHS(ICSI)/UNEP/UNESCO 1989) correspond to a sea level rise in the order of one millimetre, or even less. However, in densely populated high mountain areas such as the European Alps glaciers are also unique resources of fresh water for agriculture and industry (e.g., Braun et al. 1999, Kuhn & Batlogg 1999, BUWAL et al. 2004), an important economic component of tourism (e.g., Abegg 1996, Elsasser & Bürki 2005) and hydro-power production (e.g., UNEP 1992, Barnett et al. 2005), and a potential source of serious natural hazards (e.g., Kääb 1996, Huggel 2004, Salzmann et al. 2004, Noetzli et al. in press, Fischer et al. subm.).

Glaciers react in a highly sensitive manner to climate changes due to their proximity to the melting point. They reflect secular warming at a high rate and global scale (Haeberli 2004). A worldwide collection of information about ongoing glacier changes was initiated in 1894 when the International Glacier Commission at the Sixth International Geological Congress was founded in Zurich, Switzerland. It was hoped that long-term glacier observations would give insight into global uniformity and terrestrial or extra-terrestrial forcing of past, ongoing and future climate, and processes of climatic change such as the formation of ice ages (Forel 1895, Haeberli in press). Today, glaciers are considered key indicators within global climate-related observing systems for early detection of trends potentially related to the anthropogenic greenhouse effect (IPCC

2001, WMO 2003, Haeberli 2004). Glaciers are not only an important climate proxy but they also have become a leading icon in the present-day debate on climate change.

Numerous studies investigated the relationship between glaciers and climate change based on reconstructions (e.g., Zumbühl 1980, Nicolussi 1994, Nicolussi & Patzelt 2000, Holzhauser et al. 2005), in-situ measurements (e.g., Kuhn et al. 1985, Hagen et al. 1998, Hoelzle et al. 2003) and/or remote sensing data (e.g., Bayr et al. 1996, Maisch et al. 2000, Paul 2004, Kääb 2005), some of them in combination with statistical (e.g., Zumbühl & Messerli 1980, Braithwaite 1981, Letréquilly & Reynaud 1990), analytical (e.g., Kuhn 1981, Kuhn 1989, Oerlemans 2001, Klok & Oerlemans 2003, Oerlemans 2005) or numerical models (e.g., Greuell 1989, Oerlemans 2001, Klock & Oerlemans 2002, Leysinger & Gudmundsson 2004, Paul et al. in press). Nevertheless, glacier-climate studies focusing on an entire mountain range and integrating in-situ measurements, remote sensing data and modelling approaches – as proposed by modern monitoring strategies (cf. Haeberli 2004) – are largely lacking.

1.2 Research objectives

This study aims to investigate glacier fluctuations in the European Alps after 1850. For this purpose, glacier inventories, in-situ measurements and numerical models are used to analyse past, present and future glacierisation of the entire Alpine mountain range. The main tasks are:

- the compilation of all available glacier data in the European Alps;
- the spatio-temporal analysis of glacier fluctuations after 1850; with
- a special focus on glacier mass balance of the extraordinary year 2003;
- the comparison of glacier changes in the Alps with other mountain ranges;
- the investigation of Alpine glacier sensitivity to climate change (precipitation and temperature); and
- the assessment of impacts of climate change scenarios on Alpine glacierisation.

1.3 Outline of the thesis

The thesis is divided into three parts (see Figure 1.1):

Part A provides an overview of the present research and consists of seven major chapters: this introduction constitutes *Chapter 1*. *Chapter 2* gives a general overview of the geographical location of the European Alps. *Chapter 3* summarises the thematic and scientific background on which this study is based. It briefly describes the climate change in Europe from the Last Glacial Maximum (LGM, about 21,000 y BP) until present and climate scenarios for the 21st century. Further, an introduction is given of the cryosphere model, the glacier energy and mass balance as well as of the basic process chain between a climatic change and glacier fluctuations, followed by a review of the current state-of-the-art in modelling the glacier response to a change in climate, a summary of international glacier monitoring and, finally, an outline of Alpine glacier fluctuations before 1850. *Chapter 4* consists of the summaries of the five papers which constitute the core of this thesis. A general discussion of the main tasks and findings is given in *Chapter 5*. Conclusions and outlook are found in *Chapters 6* and *7*, respectively.

Part B contains a full version of the five papers which comprise the main scientific work.

Part C is an appendix of scripts for ArcGIS 9.x from the geographical information systems (GIS) vendor ESRI, developed for, and applied in, *Papers IV* and *V* of this study.

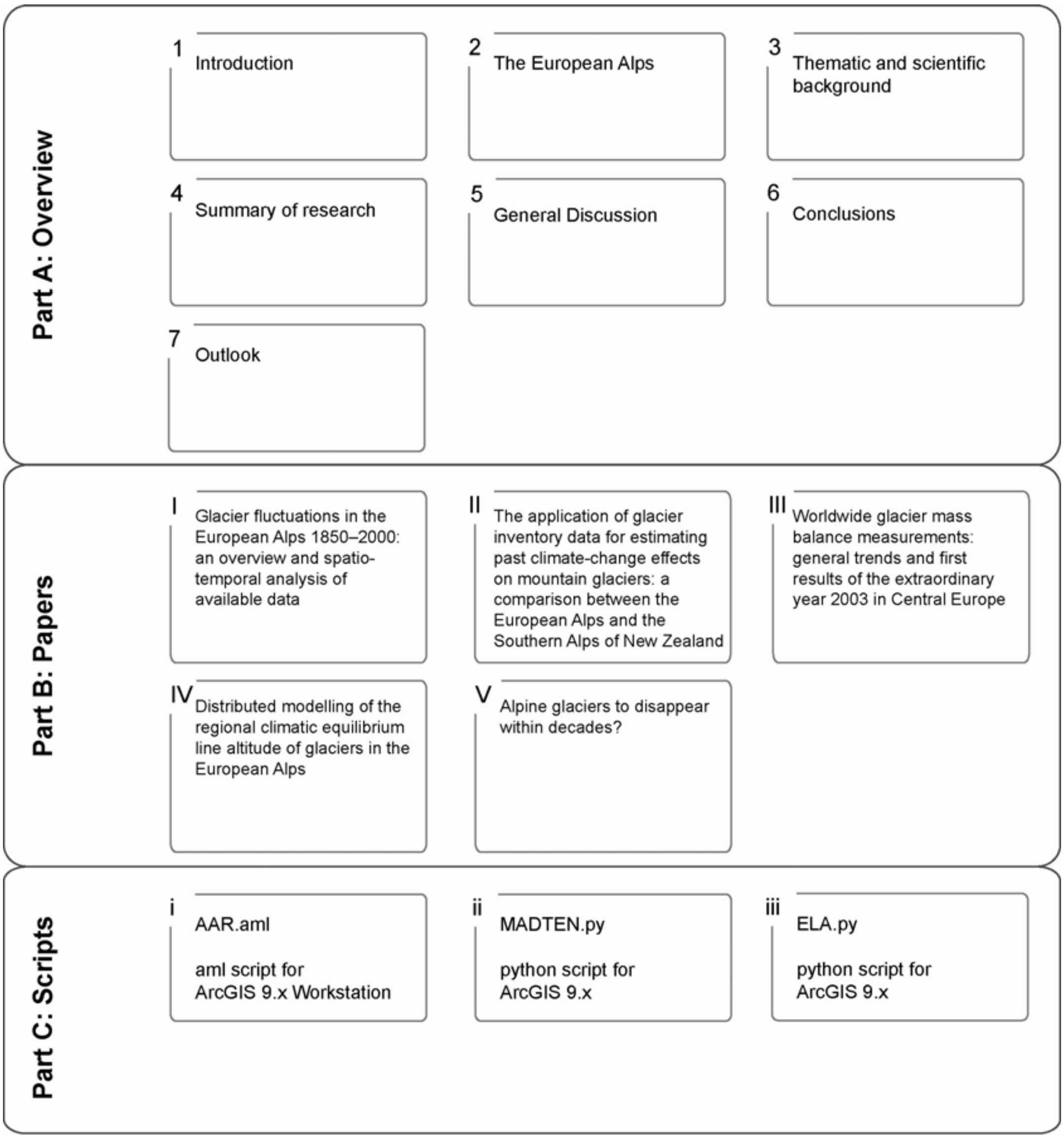

Figure 1.1: Schematic structure of the thesis, showing the contents of the three main parts Overview, Papers and Scripts.

2 The European Alps

The European Alps are located between 5–16° E and 43–48° N (Figure 2.1) and range from the Mediterranean sea level to a maximum altitude of 4,807 m a.s.l. at Mont Blanc summit in the western part of the Alps. The most eastern summits above 4,000/3,000/2,000 m a.s.l. are the Piz Bernina (4,052 m a.s.l.) in the Engadine at the Swiss-Italian border, Hochalmspitze (3,355 m a.s.l.) in the Upper Tauern region and the Schneeberg (2,075 m a.s.l.) in Austria (Bachmann 1978). The Alps are arcuate in form at their western end where they extend some 250 km north-south. Their orientation is almost west-east through Switzerland, where the ranges are less than 100 km in total width, and through Austria where they broaden to 150 km, but subdivided by pronounced longitudinal valleys such as the Inn and the Drau (Barry 1992).

In general, the climate of the Alpine region is characterised by a high degree of complexity, due to the interactions between the mountains and the general circulation of the atmosphere, which result in features such as gravity wave breaking, blocking highs, and Föhn winds. A further cause of complexity inherent to the Alps results from the competing influences of a number of different climatological regimes in the region, namely Mediterranean, Continental, Atlantic, and Polar (Beniston 2005a).

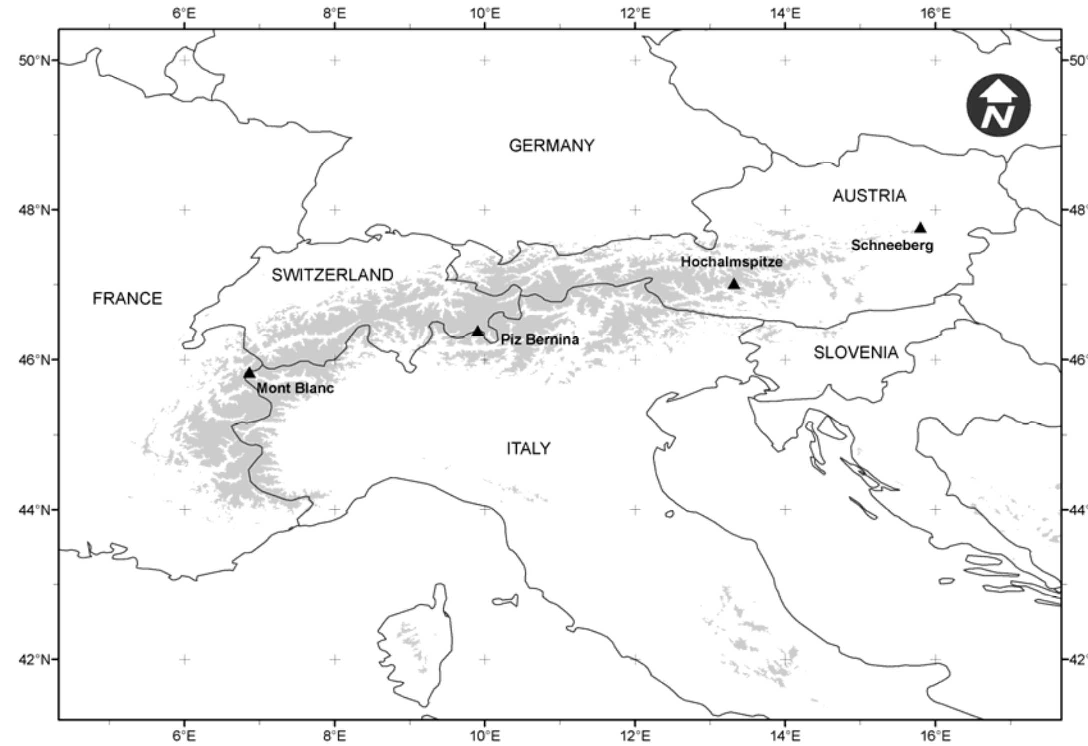

Figure 2.1: Greater Alpine Region with the highest peak (Mont Blanc) and the most eastern summits above 4,000/3,000/2,000 m a.s.l. Background: European HYDRO1k-DEM, elevations above 1,500 m a.s.l. are shaded in grey. Source: Land Processes Distributed Active Archive Center (LP DAAC), located at the U.S. Geological Survey (USGS) Centre for Earth Resources Observation and Science (EROS), http://LPDAAC.usgs.gov. Country boundaries provided by ESRI.

3 Thematic and scientific background

This chapter summarises the thematic background needed to understand the relevance and context of the present study, and gives an overview of the state-of-the-art of climate-glacier research on which this study is based.

3.1 Climate

Climate varies on all time scales, from years to decades to millennia to millions and billions of years. This climate variability can arise from a number of factors, some external to, and others internal to, the system (Bradley 1999, Ruddiman 2001). *Chapters 3.1.1* to *3.1.3* give a brief overview of Lateglacial and Holocene climate (based on Rahmstorf 2002, Jones & Mann 2004 and Wakonigg 2005), as this period is the most relevant to assessing the potential uniqueness of recent climate changes and the projected changes in the 21st century. The latter are summarised in *Chapter 3.1.4*.

3.1.1 Pleistocene and Lateglacial

The Pleistocene climate epoch was marked by major glacial/interglacial oscillations in global ice volume and ended with the most recent glacial period culminating about 21,000 y BP (the LGM), when continental ice sheets extended well into the mid-latitudes of North America and Europe. Global annual mean temperatures were probably about 4 °C colder than during the mid-20th century (Jones & Mann 2004), with larger cooling at higher latitudes and little or no cooling over large parts of the tropical oceans (Ruddiman 2001). In the European Alps, summer temperatures were depressed by more than 10 °C (Patzelt 1980) and precipitation reduced by at least 50–70% (Haeberli & Penz 1985) as compared to current values. Deglaciation during the transition to the Holocene happened not continuously but with several intermittent rebounds to colder climate. According to Rahmstorf (2002), three key factors need to be considered for this complex sequence of events: the changes in insolation (due to the Milankovich cycles) which must have initiated deglaciation, the rise in atmospheric CO_2 levels providing a strong global warming feedback, and changes in ocean circulation. An explanation for the cooling events can be found in the domination of the northern sites by the state of the Atlantic thermohaline circulation. The North Atlantic Deep Water formation in the Nordic Sea was weakened or even stopped because of the high meltwater influx from the shrinking ice sheets. The last major cooling period happened about 12,000 y BP and is known as the Younger Dryas (YD; Stocker 1996, Rahmstorf 2002). A final northern cooling in the history of deglaciation was a short event occurring 8,200 y BP, which has also been linked to a meltwater-induced weakening of the thermohaline circulation (Renssen et al. 2001). Since then, climate has been relatively warm and stable, but, nonetheless with significant variability on centennial and millennial time scales. A schematic Lateglacial and Holocene summer temperature curve for the European Alps, derived from pollen analysis and named after glacier states, is presented by Patzelt (1980) and shown in Figure 3.1. Here, The YD can be allocated to the Egesen glacier advance with prominent moraines all over the Alps (e.g., Mayr & Heuberger 1968, Kerschner 1978, Furrer et al. 1987).

3.1.2 Holocene

During the early millennia of the Holocene, atmospheric circulation, surface temperature, and precipitation patterns appear to have been substantially different from the present day, with evidence, for example, of ancient lakes in what is now the Sahara Desert (Street & Grove 1979). During the so-called mid-Holocene Optimum of 5,000–6,000 y BP, surface temperatures appear to have been milder in some parts of the globe, particularly in the extratropics of the Northern Hemisphere during summer (Webb & Wigley 1985). Recent modelling studies suggest that mid-Holocene surface temperatures for annual and global means may actually have been cooler than those of the mid-20th century, even though extratropical summers were likely somewhat warmer (Kitoh & Murakami 2002). Extratropical summer temperatures appear to have cooled over the subsequent four millennia. Orbital influences likely influenced large-scale climate over this time frame through an interaction with monsoon and El Niño-Southern Oscillation phenomena (Jones & Mann 2004). For Europe, Wakonigg (2005) describes in this period, sometimes referred to as the Neoglacial because of punctuated glacier advances (and retreats), three subperiods of warmer temperatures (Optimum) and three periods with colder temperatures (Pessimum), with reference to the 20th century mean (cf. Figure 3.1): a main Pessimum around 2,000–2,500 y BP, the Roman Optimum around 2,000 y BP, the Pessimum of Migration around 1,500 y BP, the Mediaeval Warm Period from around (AD) 800 to the onset of the Little Ice Age (LIA) around 1300, which lasted until the mid-19th century and was followed by the recent warming.

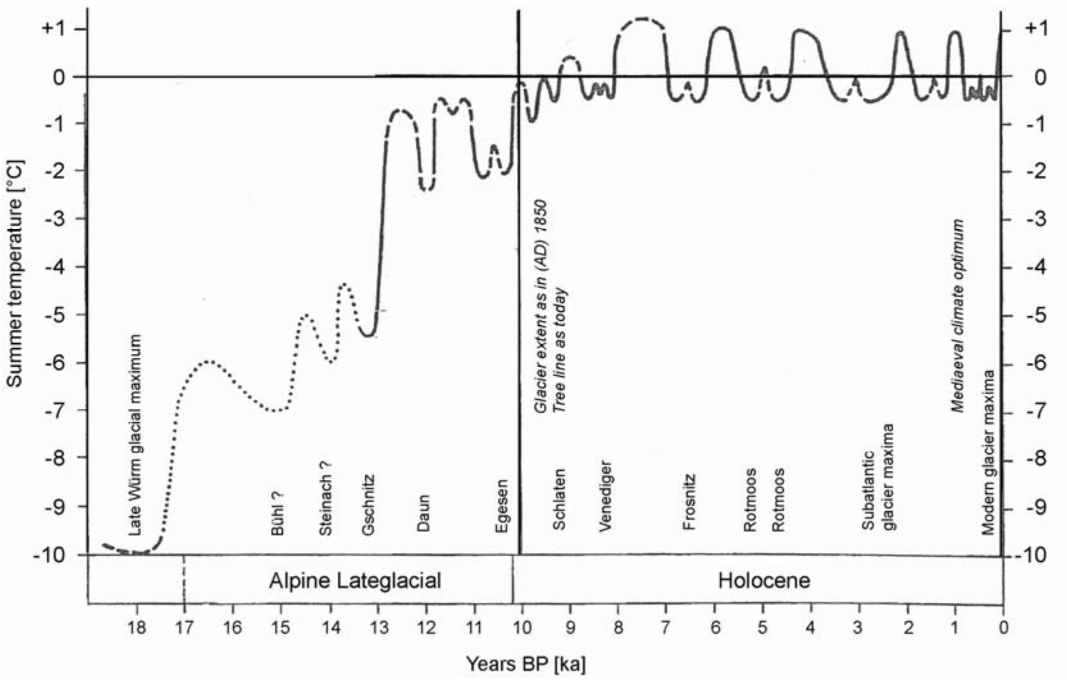

Figure 3.1: Alpine summer temperatures of the Lateglacial and the Holocene, as estimated from tree line and snow line variations. Slightly modified after Patzelt (1980).

The above described subperiods and also Figure 3.1 are somewhat simplified and climate trends become more complex when analysed with increasing spatial and temporal resolution. Towards the present time, there is an increase in the availability and spatial density of climate proxy data and more detailed studies become available (see below). Recent developments in climate research lead to gridded climatologies with high temporal (monthly, seasonal or annual) and spatial (100 m to 60 km) resolutions (e.g., Schwarb 2000, Luterbacher et al. 2004, New et al. 1999, 2000, Efthymiadis et al. 2006).

3.1.3 The last millennium

Variations in the Northern Hemisphere's surface temperature over the last millennium are shown in Figure 3.2, as published by IPCC (2001). The temperature variations have been reconstructed from proxy data (tree rings, corals, ice cores and historical records) calibrated against thermometer data. A general negative temperature trend from 1000 until the mid-19th century followed by a distinctive warming of 0.6 ± 0.2 °C over the last 100 years can be seen. Jones & Mann (2004) provide a review of the evidence for climate change over the past several millennia from instrumental and high-resolution climate proxy data sources and climate modelling studies. They conclude that the 20th century warmth is unprecedented at hemispheric and, likely, global scale and that natural factors appear to explain relatively well the major surface temperature changes of the past millennium through the 19th century. They also hold that only anthropogenic forcing of climate can explain the recent anomalous warming in the late 20th century.

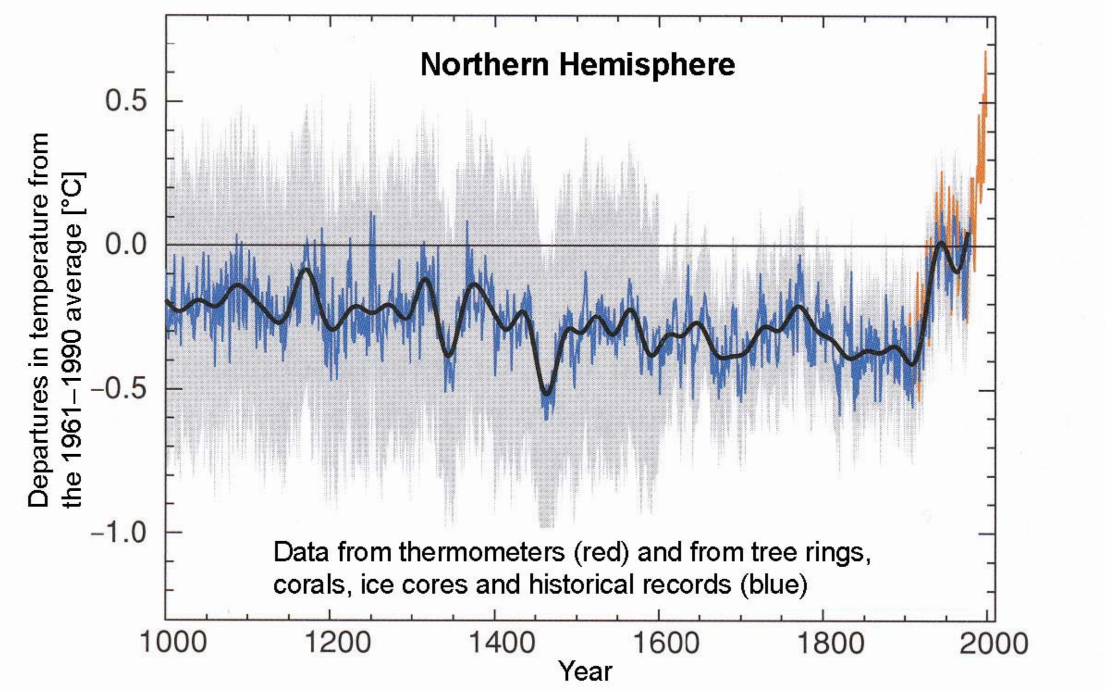

Figure 3.2: Variations in the Northern Hemisphere's surface temperature over the last millennium. The year by year (blue curve) and 50-year average (black curve) variations in the surface temperature for the past 1,000 years have been reconstructed from proxy data calibrated against thermometer data (red curve). The 95% confidence range in the annual data is represented by the grey region. Source: IPCC (2001).

Casty et al. (2005) reconstructed Alpine temperature and precipitation since 1500 and compared it with smoothed regional (Luterbacher et al. 2004, Pauling et al. 2006), large-scale (Jones et al. 1998) and homogenised instrumental (Böhm et al. 2001, Auer et al. 2005) data (Figure 3.3). They show that the global and Northern Hemispheric warming at the end of the 20th century (e.g., Jones et al. 1998, IPCC 2001) and in Europe (e.g. Luterbacher et al. 2004) is also prominent in the greater Alpine region and occurred in two main stages: between 1880 and 1945 and since 1975. The years 1994, 2000, 2002 and particularly 2003 were the warmest since 1500 in the greater Alpine region, with 2003 being 1.7 °C warmer than the mean of the 20th century (see also Schär et al. 2004). In addition, a period of warmth around 1800 and periods of distinct cooling roughly around 1600, 1700 and 1900 are seen in all records. Pfister (1992, 1999) and Luterbacher et al. (2004) found similar cooling periods for Europe. The Alpine temperature variations show larger amplitudes compared to the one averaged for the Northern Hemisphere, confirming earlier results, for instance from Luterbacher et al. (2004), Jones & Mann (2004) and Beniston (2005a). In contradiction to the expectation for increased precipitation due to increased temperature (cf. IPCC 2001), the Alpine precipitation time series do not indicate significant trends (Casty et al. 2005).

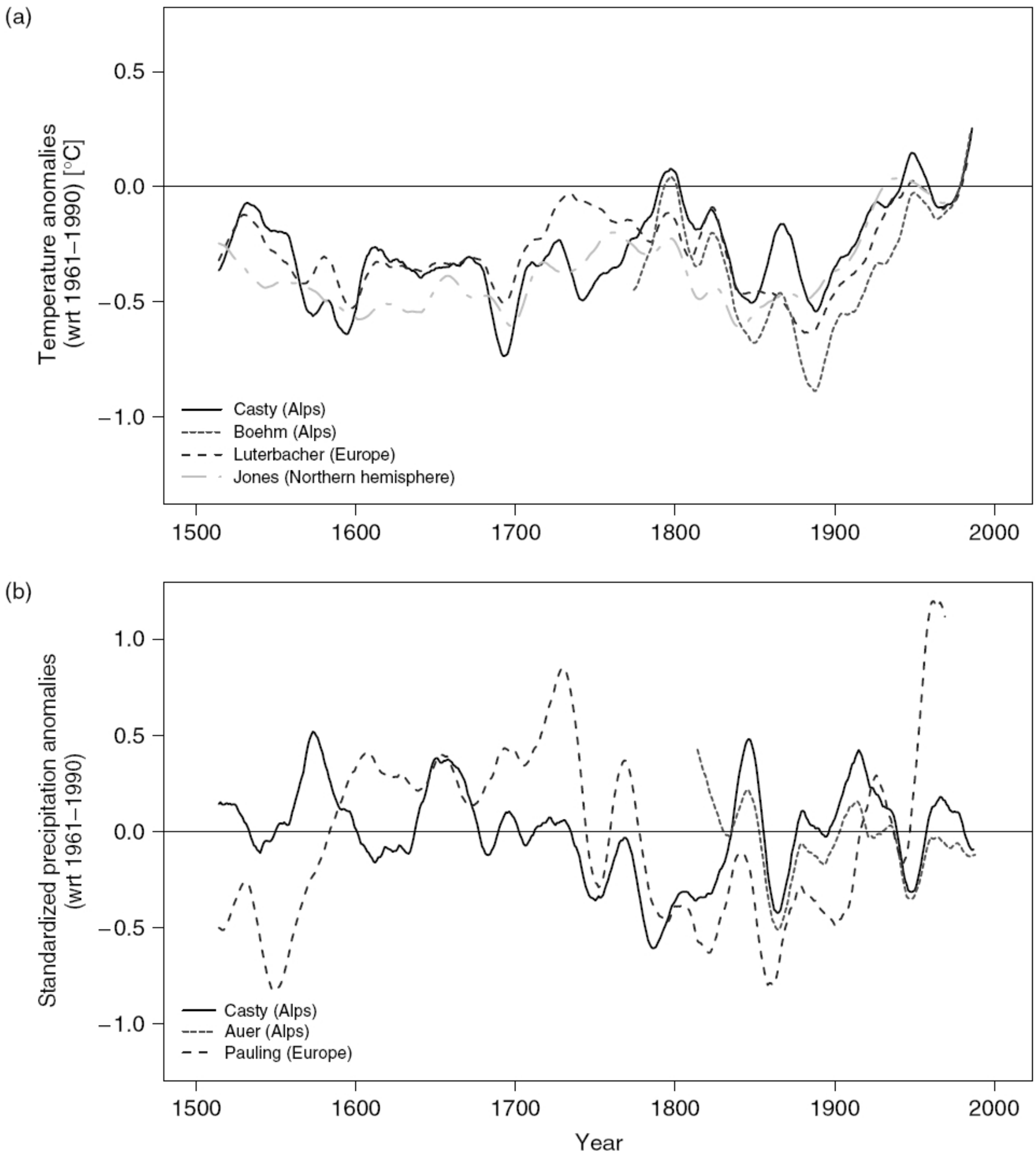

Figure 3.3: Reconstructed temperature and precipitation series since 1500. (a) Smoothed (31-year triangular filter) time series of temperature reconstructions for the Northern Hemisphere (Jones et al. 1998), Europe (Luterbacher et al. 2004), and the Alpine region (Böhm et al. 2001 and Casty et al. 2005). The temperature anomalies are calculated with regard to the 1961–1990 mean. Note that Jones et al. (1998) is published as summer temperature reconstruction. (b): same as (a) but for standardised annual precipitation anomalies (reference period 1961–1990) for Europe (Pauling et al., 2006) and the Alpine region (Auer et al. 2005 and Casty et al. 2005). Figure after Casty et al. 2005.

3.1.4 Future scenarios

Current climate models are able to simulate quantitatively the observed warming and to assign different causes to the climate changes observed over the last centuries. Until about 1950 the mean global temperature was controlled mainly by natural variability in net radiation due to cycles of solar irradiative forcing and volcanic eruptions, whereas the temperature rise after 1950 can only be explained when including the anthropogenic increase in greenhouse gases (IPCC 2001). For the IPCC special report on emissions scenarios (IPCC 2000), 40 scenarios were developed to describe the emissions from important greenhouse gases and aerosols between 1990 and 2100, based on different plausible assumptions of economic and population growth, hypotheses related to technological advances, the rate at which the energy sector may reduce its dependency on fossil fuels, and other socio-economic projections related to deforestation and land-use changes (Figure 3.4). Based on the projected levels of emissions from greenhouse gases and aerosols in the atmosphere (Figure 3.5a), global and regional climate models are used to provide insight into the potential future climate. According to these scenarios, the increase in global mean temperature by the end of the 21st century ranges from 1.4 °C to 5.8 °C (Figure 3.5b).

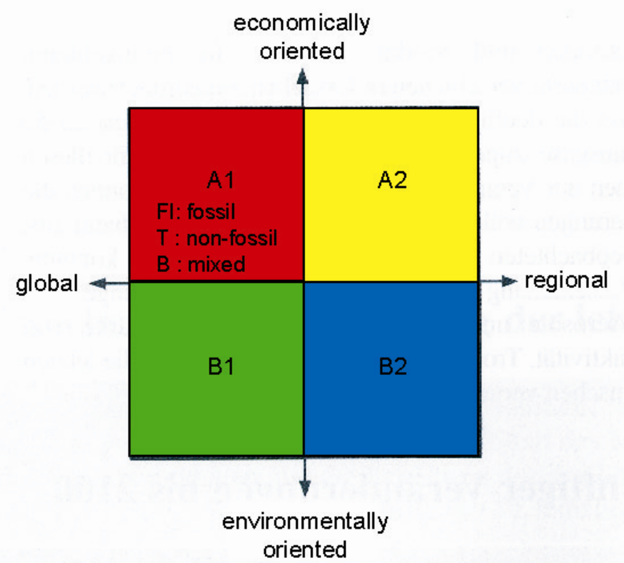

Figure 3.4: The four basic types of IPCC emission scenarios for the 21st century. Type A1 is represented by three scenarios (FI, T, B) using different energy systems. Source: IPCC (2001).

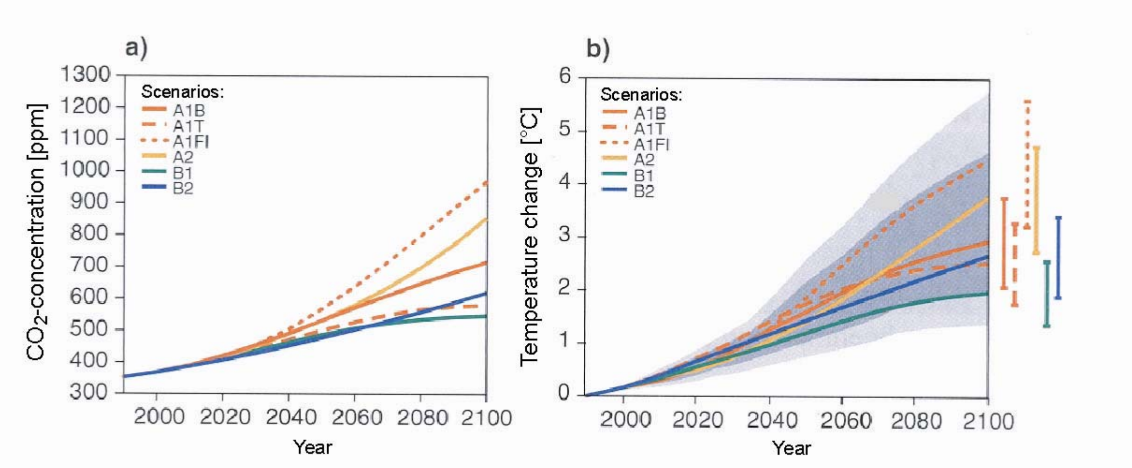

Figure 3.5: IPCC-scenarios of atmospheric CO_2-concentration (a) and corresponding mean global temperature change (b) between 1990 and 2100. The projected temperature rise at the end of the 21st century ranges between 1.4 and 5.8 °C. The region of the temperature change of all scenarios and all model runs is shown in light grey, the region of the average of all model runs for all scenarios is shown in dark grey. Source: IPCC (2001).

Within the EU-funded PRUDENCE project (cf. Christensen et al. 2002), a series of regional climate models has been applied to the investigation of climatic change over Europe for the last thirty years of the 21st century, enabling changes in a number of key climate variables to be assessed. The regional models operating at a 50 km resolution have simulated two thirty-year periods: the 'current climate' or the 'control simulation' for the period 1961–1990, and the future 'greenhouse gas climate' for the period 2071–2100. These simulations suggest that the Alpine climate in the late 21st century will be characterised by warmer and more humid winter conditions, and much warmer and drier conditions in summer (Beniston 2005b). Frei (2004) presented probabilistic climate projections for mean temperature and mean precipitation for Switzerland, based on the climate simulations of the PRUDENCE project. From 1990–2050, he expects winter temperature to rise by 1.8 °C (range: 0.9–3.4 °C) and summer temperature by 2.7 °C (range: 1.4–4.9 °C) for the northern side of the Alps, with only minor differences for southern Switzerland (Figure 3.6). For winter precipitation he predicts for the same period (1990–2050) an increase of 8% (11%) and for summer precipitation a decrease of 17% (19%) for the northern (southern) side of the Swiss Alps (Figure 3.7).

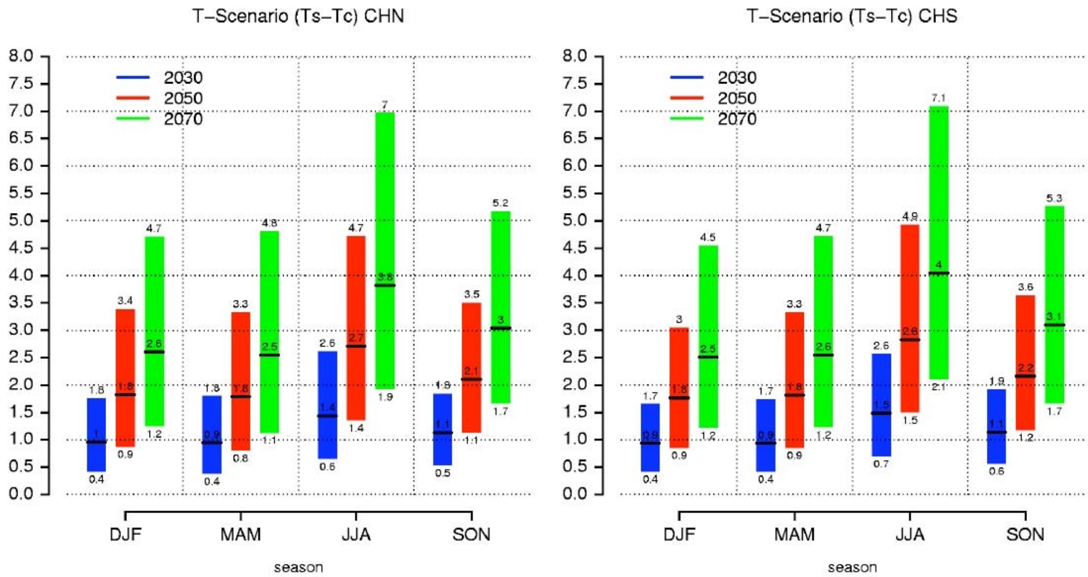

Figure 3.6: Probabilistic temperature projections for northern (left) and southern (right) Switzerland. Changes in the mean temperature in degree Celsius (y-axis) from 1990 to 2030 (blue), to 2050 (red) and to 2070 (green) are shown with the median (black line) and the 95% confidence interval (coloured bars) for the four seasons (DJF, MAM, JJA and SON). Figure from Frei (2004).

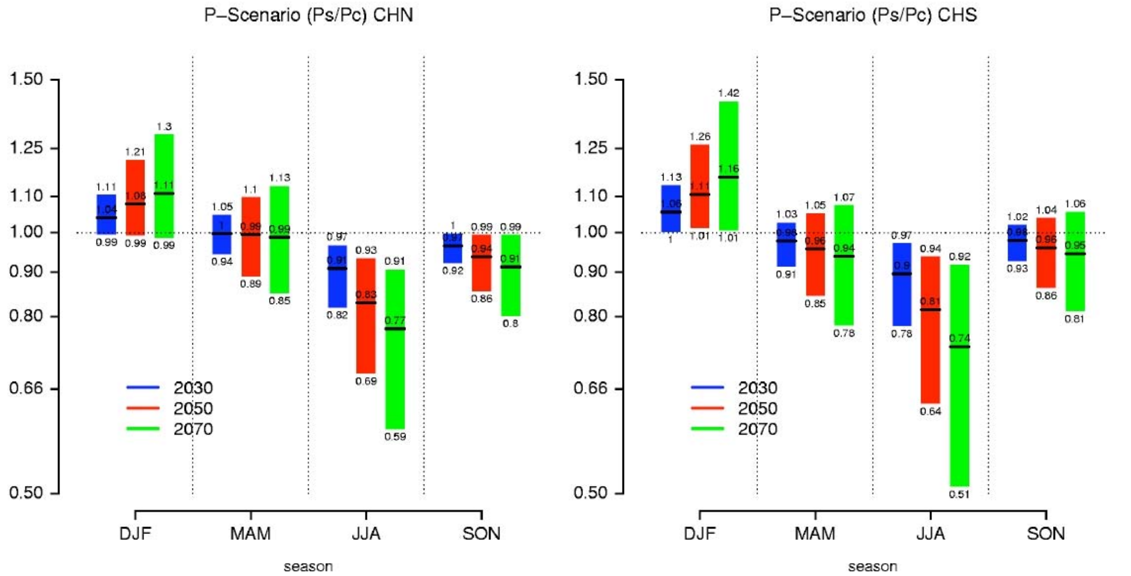

Figure 3.7: Probabilistic precipitation projections for northern (left) and southern (right) Switzerland. Changes in the mean seasonal precipitation (y-axis) are shown as ratios between future (2030, 2050 and 2070) and present (1990) state, with the median (black line) and the 95% confidence interval (coloured bars) for the four seasons (DJF, MAM, JJA and SON). Figure from Frei (2004).

3.2 Glaciers and climate change

This chapter briefly introduces the so-called cryosphere model, i.e., the concept explaining glacier distribution as a function of air temperature and precipitation, the equations of the energy balance at the glacier surface and of the glacier mass balance, as well as the principle process chain between a climatic change and the glacier reaction. A corresponding parameterisation scheme by Haeberli & Hoelzle (1995) is presented as it provides the basis for the approach used in *Paper II*. In addition, a review of modelling approaches to determine the response of glaciers to a change in climate is given, followed by an overview of international glacier monitoring and its strategies. Finally, Alpine glacier fluctuations between the LGM (about 21,000 y BP) and the end of the LIA (around 1850) as well as corresponding reconstruction methods are presented briefly.

3.2.1 The cryosphere model

Glaciers form where the winter accumulation does not melt entirely during summer, at least in most years. This perennial snow gradually densifies and transforms through different processes of snow metamorphosis (mutual displacement of crystals, changes in size and shape and internal deformation) into firn and finally, after the interconnecting air passages between the grains are closed off, into ice (Paterson 1994). Glacier distribution is primarily a function of mean annual air temperature and annual precipitation (e.g., Shumsky 1964, Kuhn 1981, Oerlemans 2001), tending to be located in wet and cold places. Wet implies a region not too far from a moisture source (the ocean) located in upwind direction. Cold implies high latitude or altitude. Thus, temperature and precipitation are not entirely independent factors. Towards the poles, mean precipitation becomes less and less, because of the decreasing moisture-holding capacity of the air with lower temperature. As a result, the windward side of the high mountain ranges located in the westerlies is the best place to find glaciers. Indeed, the glaciers with the highest mass turnover are to be found, for instance in Norway, Alaska, Patagonia and New Zealand (Oerlemans 2001). The mountain ranges with persistent atmospheric circulation show a marked asymmetry in the precipitation regime. The Alps present another picture, as the storm tracks move in from different directions over the year. Here a pattern of high precipitation in the outer mountain ranges and somewhat drier conditions in the interior can be found (Schwarb 2000).

Driven by gravity, the cumulating snow/ice mass starts to move by viscous creep (Nye 1952, Glen 1955) and basal sliding (Lliboutry 1971) until the glacier front reaches a position such that the net mass balance over the entire glacier equals zero. Thereby, the theoretical line on the glacier that defines the altitude at which annual accumulation equals ablation is called the equilibrium line altitude (ELA). It represents the lowest boundary of climatic glacierisation (Paterson 1994). The region above the ELA shows a net mass gain over the year and is thus called the accumulation area, whereas the region below the ELA is called the ablation area with annual net mass loss, normally increasing towards the glacier terminus (Figure 3.8).

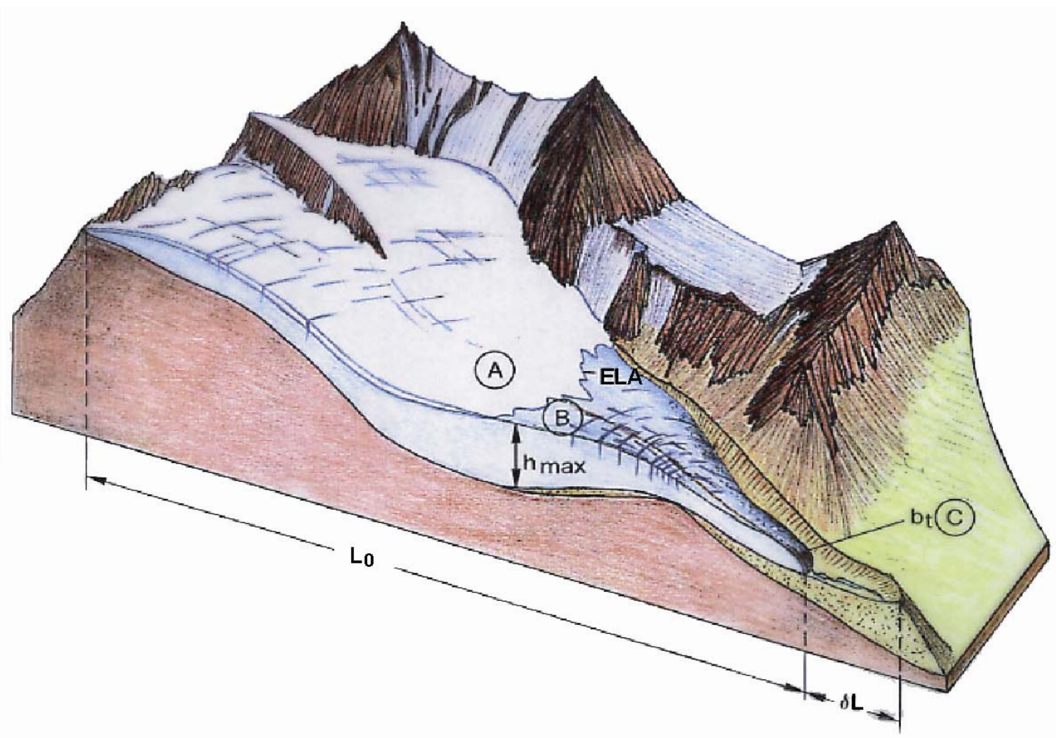

Figure 3.8: Schematic plot of a mountain glacier. Glacier elements, such as accumulation area (A), ablation area (B), glacier terminus (C), ELA, maximum ice thickness (h_{max}), glacier length (L_0), length change (δL) and ablation at the glacier terminus (b_t), are explained in the text. Figure from Haeberli (1995).

Haeberli (1983) presented a schematic diagram of glacier limits, the so-called cryosphere model (Figure 3.9), after Shumsky (1964), describing the distribution of glaciers primarily as a function of mean annual air temperature and annual precipitation. He notes that, as a general rule, in humid-maritime regions the ELA is at low altitude because of the great amount of ablation required to eliminate the deep snowfall. Temperate glaciers dominate these landscapes. Such ice bodies, with relatively rapid flow, exhibit a high mass turnover and react strongly to atmospheric warming by enhanced melt and runoff. Under dry continental conditions the ELA is at high elevation. In such regions, glaciers are polythermal or cold and feature a low mass turnover.

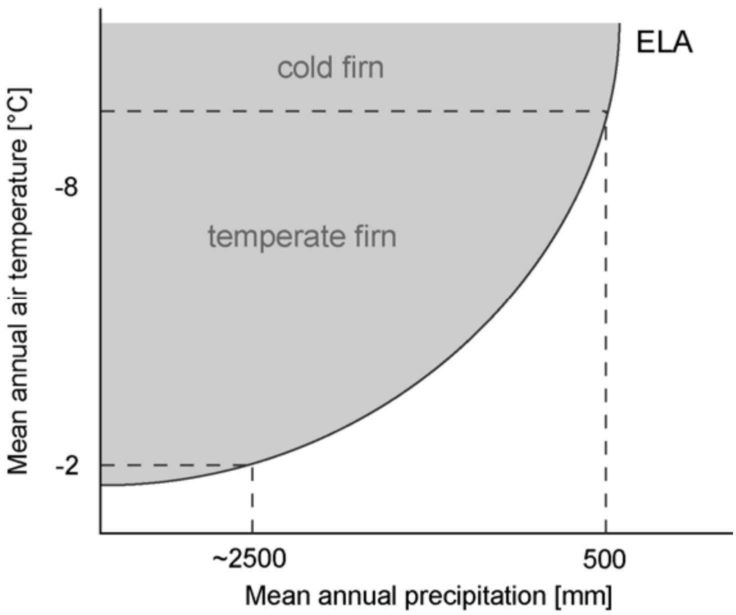

Figure 3.9: Schematic diagram of glacier limits as a function of mean annual air temperature and average annual precipitation. Figure modified after Shumsky (1964) and Haeberli et al. (1989a).

3.2.2 Energy balance at the glacier surface

At its surface, the glacier interacts with the atmosphere. Mass and energy fluxes at the glacier surface affect mass gain or loss of the entire glacier, and subsequently drive the ice flow. The energy balance at the glacier surface governs the total amount of energy E available for melting, and can be expressed according to Oerlemans (2001):

$$E = Q(1-\alpha) + L_{in} + L_{out} + H_S + H_L + G \qquad (1)$$

where Q is the global radiation, α the albedo, L_{in} and L_{out} the incoming and outgoing long-wave radiation, respectively. H_S is the turbulent sensible heat flux, H_L the turbulent latent heat flux, and G the heat flux into or out of the glacier's snow, firn and ice due to conduction and convection, and due to penetration of solar radiation. Positive values of E are defined as fluxes towards the glacier surface. The largest fluxes are the radiative fluxes, typically a few hundred watts per square metre. A large part of the solar radiation reaching the glacier surface is reflected, depending on the albedo (most in the case of fresh snow, less in the case of old snow or bare ice and very little when the surface is covered by morainic material). For long-wave radiation the glacier surface is very dense and, hence, the amount of long-wave radiation reflected by the surface is negligible. Incoming and outgoing long-wave radiation compensate each other to a large degree. The effects of clouds on the short-wave and long-wave radiation budgets are of opposite sign. More clouds imply less short-wave radiation and more long-wave radiation. The net effect depends on surface albedo and cloud transmissivity. Turbulent exchange of heat and moisture can be quite significant, especially in winter (when the sun is low) or in summer when air temperature is high. The fluxes are in the direction of the gradients in the mean temperature and humidity profiles. When air temperature is above freezing, the sensible heat flux is always towards the surface. In such conditions the latent heat flux can go in both directions, depending on the humidity of the air. The saturation vapour pressure for a melting glacier surface is 610.8 Pa. If an air mass has a temperature of 10 °C, the gradient in vapour pressure and hence the vapour flux changes sign at a relative humidity of about 50%. At lower humidity evaporation cools the surface, at higher humidity condensation heats the surface. The temperature of precipitation may differ from the temperature at the surface. This implies that precipitation may add or remove heat from the existing snowpack or ice surface. However, the energy fluxes involved are minor. The fluxes inside, and at the base of the glacier are much smaller than those between the atmosphere and the glacier surface, except for the flow of meltwater, which represents a latent heat flux. In snow and firn, convection by air motion transports heat and water vapour. These fluxes are small but important for the metamorphosis of snow crystals. A more detailed discussion of the different components of the glacier energy balance is given, for instance, in Kuhn (1981, 1989, 1990), Funk (1985), Greuell & Oerlemans (1986), Paterson (1994), or Oerlemans (2001).

3.2.3 Glacier mass balance

Glacier mass balance b is the result of accumulation and ablation processes and writes according to Kuhn (1981) as:

$$b = c - a \tag{2}$$

Accumulation c includes all processes bringing mass into the glacier system:

$$c = P_{solid} + P_{stored} + P_{condensation} + D_{+} + D_{avalanches+} \tag{3}$$

The most important process, thereby, is the solid precipitation (P_{solid}) followed by accumulation due to snowdrift (D_{+}). Hoinkes (1957) and Kotlyakov (1973) showed that accumulation due to snowdrift in the Alps can be in the same order as that due to solid precipitation and, therefore, cannot be neglected on Alpine glaciers. Other processes are the snow redistribution due to avalanches ($D_{avalanches+}$), accumulation due to freezing rain (P_{stored}) and the formation of surface hoar ($P_{condensation}$). Superimposed ice can become an important (internal) accumulation component, mainly on poly-thermal glaciers (e.g., Paterson 1994).

Ablation a comprises all processes leading to a mass loss of the glacier:

$$a = M + S + D_{-} + D_{avalanches-} + D_{calving} \tag{4}$$

The dominant ablation process is melting of snow and ice (M), mainly depending on the energy balance of the glacier surface. Other processes are sublimation (S) (which is an important process in a dry and cold climate), snowdrift (D_{-}), snow erosion due to avalanches ($D_{avalanches-}$) and calving ($D_{calving}$). Melting at the glacier bed amounts to values of only a few millimetres per year and, thus, can be neglected in comparison with surface effects (e.g., Paterson 1994).

3.2.4 Glacier reaction to changing climate

The reaction of a temperate glacier to a climatic change, or rather to a corresponding change in mass balance, was analysed theoretically by Nye (1960). Glacier reaction to climatic change involves a complex chain of processes (energy balance – mass balance/englacial temperature – geometry/flow – length change), described by Meier (1984) or by Haeberli (1994; including englacial temperature), and can best be understood and numerically simulated for a few glaciers only, which have been studied in great detail (e.g., Greuell 1992, Schmeits & Oerlemans 1997, Zuo & Oerlemans 1997b, Oerlemans 2001, Leysinger-Vieli & Gudmundsson 2004). The complication, however, disappear if the time intervals analysed are sufficiently long, i.e., longer than it takes a glacier to complete its adjustment to a climatic change. Based on the calculations from Johannesson et al. (1989), Haeberli & Hoelzle (1995) presented the following scheme of the glacier reaction to a step change in climate and a corresponding change in ELA for temperate glaciers (Figure 3.10, cf. Haeberli 1994): An assumed step change (δ) in ELA causes an immediate step change in specific mass balance (b = total mass change divided by glacier area). The specific mass balance is thus the direct, undelayed reaction of a glacier to climatic forcing. A change in specific mass balance (δb) is calculated from the shift in ELA (δELA), the gradient of mass balance with altitude (db/dH), and the distribution of glacier surface area with altitude (hypsography). The hypsography represents the local or individual topographic part of the glacier sensitivity to a climatic change, whereas the mass balance gradient mainly reflects the regional or climatic part (Kuhn 1990). After a certain reaction time (t_r) following a change in mass balance, the length of a glacier (L_0) will start changing and finally reach a new equilibrium ($L_0 + \delta L$) after the response time (t_a). The length change is thus the indirect, delayed reaction of a glacier to climatic forcing. After full response, continuity requires, according to Nye (1960), that:

$$\delta L = \frac{L_0 \cdot \delta b}{b_t} \tag{5}$$

where b_t is the (annual) ablation at the glacier terminus. This means that, for a given change in mass balance, the length change is a function of the original length of a glacier, and that the change in mass balance of a glacier can be quantitatively inferred from the easily observed length change and from estimates of ELA and mass balance gradients. The response time (t_a) of a glacier is related to the ratio between its maximum thickness (h_{max}) and its ablation at the terminus:

$$t_a = \frac{h_{\max}}{b_t} \tag{6}$$

In contradiction to earlier theoretical analysis (e.g., Nye 1963, Paterson 1994), corresponding values for Alpine glaciers are found to be typically several decades to slightly less than a century. Aletsch Glacier as the largest Alpine glacier, for instance, with h_{max} close to 900 m and b_t around 12 m per year, has an estimated response time of some 70 to 80 years. During the response time, the mass balance b will adjust to zero again so that the average mass balance <*b*> is close to 0.5 times δb. The secular mass change of glaciers estimated in this way (<b>) can be compared directly with the few measured long-term mass balance series existing in the Alps (cf. IUGG(CCS)/UNEP/UNESCO/WMO 2005).

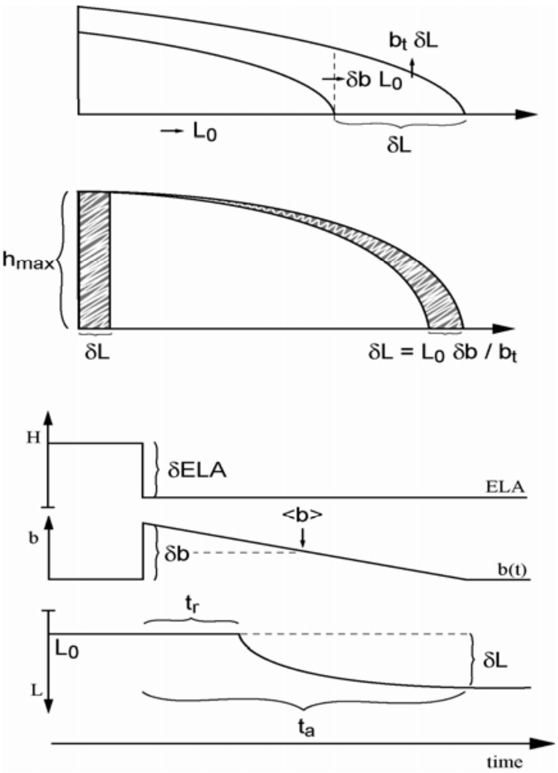

Figure 3.10: Schematic diagrams of the glacier reaction to a step change in ELA and mass balance as a function of time. Explanation is given in the text. Figure modified after Haeberli (1991).

3.2.5 Modelling approaches

Over the past decades, a whole set of different models at various levels of sophistication and spatio-temporal scales have been developed by the glacier research community (cf. Oerlemans 2001 and Hoelzle et al. 2005 for broad overviews). In general, there is a relationship between model complexity and process understanding. Simple models have the advantage of needing only a small amount of input data and thus can be applied over large regions and/or long time scales. In contrast, more sophisticated models with a higher degree of complexity allow a deeper insight into the physical processes of glacier behaviour. However, due to the large amount of required input data, they are applicable only on short time scales and to individual glaciers or small glacierised catchments.

According to Kuhn (1993), the methods/models to determine the response of glaciers to a change in climate parameters can be classified into four groups: (a) the analogy, i.e., the search for similar events in the past, (b) multivariate analysis of measured data, (c) deterministic models with explicit treatment of processes linking causes and effects, and (d) the inclusion of changes in prevalent synoptic patterns.

The principle of the analogue methods builds on the concept that the same cause leads to the same effect, but in practice the concession is made that similar causes will lead to similar effects. These types of methods are often used in palaeoglaciology where the amount of available data is often minimal. One prominent example of this type of study is the work of Gross et al. (1978). They suggested a steady-state accumulation area ratio (AAR_0) of 0.67 for glaciers in the Alps as an approximation of the steady-state ELA (ELA_0) and estimated the snowline depression of Late-Würm (20,000–10,000 y BP) glacier re-advances. With the AAR_0 method, the ELA_0 of past glaciers is fairly easy to determine from topographical glacier maps and hence often applied in palaeoclimatic studies (e.g., Kerschner 1985, Maisch 1992, Kerschner et al. 2000, Maisch et al. 2000). Another group of studies tries to relate outstanding periods (usually based on temperature and/or precipitation) to marked glacier advances and retreats (e.g. Pfister 1992, Matulla et al. 2005, Wanner et al. 2005).

The methods of multivariate analysis statistically link the observed variance of one dependent variable to the variance of one or several independent variables that are physically relevant to the problem. The understanding of the physical process involved is important for the selection of variables, but beyond that, the method in its pure form is a statistical one. Often the correlation is analysed between glacier front variation, mass balance or ice ablation on a glacier (as dependent variable) and other parameters, such as (seasonal) air temperature and precipitation series from a weather station, sunshine duration, vertical distance of atmospheric pressure levels as an indicator of free air temperature, or glacier-fed river discharge (e.g., Dreiseitl 1976, Hoinkes 1967, Lliboutry 1974, Hoinkes & Steinacker 1975, Braithwaite 1981, Letréguilly & Reynaud 1990, Kuhn et al. 1997, Schöner et al. 2000). One particular form of this method is the so-called degree-day model, which uses the high correlation of temperature with various energy balance terms (e.g., atmospheric humidity, cloudiness, incoming long-wave radiation, global radiation; cf. Kuhn 1990, Ohmura 2001). They relate glacier ablation to the product of a degree-day factor times the sum of a series of daily mean temperatures in degree Celsius (e.g., Braithwaite 1981, 1984, Braun 1985, Braithwaite & Zhang 2000, Vincent 2002, Hock 2003). Multivariate regression models, relating the change in ELA_0 to the changes of climate parameters, go back to the works of Ahlmann (1924) and Krenke (1975). More recent studies were presented by Ohmura et al. (1992), Kerschner (1996), Greene et al. (1999), Lie et al. (2003) and Shea et al. (2004). Steiner et al. (2005) applied a nonlinear back propagation neural network to reconstruct mass balance of Great Aletsch Glacier driven by seasonally resolved temperature and precipitation reconstructions.

Deterministic models explicitly treat the processes linking causes and effects. The parameterisation scheme by Haeberli & Hoelzle (1995, described in *Chapter 3.2.4*) based on Johannesson et al. (1989) uses simple algorithms and basic inventory data (glacier length, area, minimum and maximum altitude) to estimate glaciological characteristics and simulate potential climate change effects on a large number of inventoried mountain glaciers. This approach was applied by Hoelzle et al. (2003) to estimate mean specific mass balances from long-term measurements of cumulative glacier length for different mountain regions as well as by Haeberli & Holzhauser (2003) to calculate average mass balance changes from 2,000 years of reconstructed fluctuations of the Great Aletsch Glacier. Another simple analytical model is the one by Klok & Oerlemans (2003) who derived historical ELA from glacier length records by inverse modelling. Similar approaches of inverse modelling were used to derive a temperature signal from glacier front variations (Klok & Oerlemans 2004, Oerlemans 2005). The first models that analytically formulate the processes governing the exchange of heat and mass between glacier surface and atmosphere were implemented in the early 1980s by Kuhn (e.g., 1981, 1989) followed by many others (e.g., Oerlemans 1991, 2001). Other numerical models followed with explicit time dependence, which allow a better description of the processes that control the energy budget (e.g., Greuell & Oerlemans 1986, Oerlemans 1991, Oerlemans 2001). Current development goes towards spatially distributed modelling of the energy and mass balance of single glaciers or entire glacierised catchments (e.g., Arnold et al. 1996, Brock et al. 2000, Klok and Oerlemans 2002, Hock & Holmgren 2005, Paul et al. in press, Machguth et al. submitted). Other recent studies focus on the combination of mass balance and flow modelling, based either on a longitudinal profile along the glacier flowline (e.g., Greuell 1992, Paterson 1994, Schmeits & Oerlemans 1997, Zuo & Oerlemans 1997b, Van der Veen 1999, Oerlemans 2001, Leysinger-Vieli & Gudmundsson 2004) or on a spatial distributed approach (e.g., Kääb & Funk 1999, Plummer & Phillips 2002).

Another group of methods focuses on the influence of synoptic patterns (i.e., the characteristics of the pressure distribution on a scale of one to several thousand kilometres; cf. Barry & Perry 1973, Barry & Carleton 2001) on the local weather, and thus on accumulation and ablation processes. Especially in the case of the European Alps with the competing influences of different climatological regimes, a change in the prevailing synoptic patterns may strongly influence the spatial patterns of accumulation and ablation. Kerschner (2004), for instance, showed that in years with very negative mass balance values, measured at Hintereis Glacier and Sonnblick Glacier (Austria), anticyclonic weather types dominated the accumulation and especially the ablation period of the two glaciers. Schöner et al. (2000) found that the positive mass balance period between 1960 and 1980 was characterised by negative winter North Atlantic Oscillation index values, which caused an increase in the meridional circulation mode and thus a more intense north-westerly to northerly precipitation regime (cf. Reichert et al. 2001, Wanner et al. 2005). Similar studies exist for instance for Svalbard glaciers (Washington et al. 2000) or New Zealand glaciers (Fitzharris et al. 1992). Into this fourth group, one could also assign the current efforts to drive glacier models with numerical climate models (e.g., Oerlemans 2001, Reichert et al. 2001, Schneeberger et al. 2003, Weber & Oerlemans 2003) or to use glacier observations for verification of numerical climate models (e.g., Beniston et al. 1997). However, there are still some major spatial and temporal gaps between numerical climate models and (cryospheric) impact models, that would require special up-/downscaling (cf. Salzmann in preparation) and inter-/extrapolation techniques (cf. Tveito & Schöner in preparation) in order to overcome. Further efforts are currently being made to represent mountain glaciers in regional climate models on a subgrid scale to include the feedback of potential changes in the ice cover to the atmosphere and to account for their effects, such as enhanced summer runoff due to glacier melting (Kotlarski & Jacob 2005).

3.3 International glacier monitoring

The following review on international glacier monitoring and its observing strategies is from the recent issue of the ‘Fluctuations of Glaciers’ series (IUGG(CCS)/ UNEP/UNESCO 2005) published by WGMS.

Worldwide collection of information about ongoing glacier changes was initiated in 1894 with the foundation of the International Glacier Commission at the 6^{th} International Geological Congress in Zurich, Switzerland. It was hoped that long-term glacier observations would give insight into processes of climatic change such as the formation of ice ages (Forel 1895, Haeberli in press). In 1986 the WGMS started to maintain and continue the collection of information on ongoing glacier changes, when the two former services of the International Commission on Snow and Ice (ICSI), the Permanent Service on Fluctuations of Glaciers (PSFG) and the Temporary Technical Secretary for the World Glacier Inventory (TTS/WGI) were combined. Since its initiation, the goals of international glacier monitoring have evolved and multiplied. Today, the WGMS is integrated into global climate-related observation systems and collects standardised observations on changes in mass, volume, area and length of glaciers with time (glacier fluctuations), as well as statistical information on the distribution of perennial surface ice in space (glacier inventories). Thus a valuable and increasingly important data basis on glacier changes has been built up over the past century (Haeberli et al. 1989b, Haeberli 1998, 2004, in press). International assessments such as the periodical reports by the Intergovernmental Panel on Climate Change (IPCC) or the Global Climate/Terrestrial Observing System (GCOS/GTOS) plan for terrestrial climate-related observation (Cilhar et al. 1997) define mountain glaciers as one of the best natural indicators of atmospheric warming with the highest reliability ranking. The Global Terrestrial Network for Glaciers (GTN-G) of the GCOS/GTOS, aims at combining (a) in-situ observations with remotely sensed data, (b) process understanding with global coverage and (c) traditional measurements with new technologies by using an integrated and multi-level strategy (Haeberli et al. 2000). This approach, the Global Hierarchical Observing Strategy (GHOST), uses observations in a system of tiers:

Tier 1: Multi-component system observation across environmental gradients
Primary emphasis is on spatial diversity at large scales (continental-type) or along elevation belts of high mountain areas. Special attention should be given to long-term measurements. Some of the glaciers already observed (for instance, those in the American Cordilleras or in a profile from the Pyrenees through the Alps and Scandinavia to Svalbard) could later form part of Tier 1 observations along large-scale transects.

Tier 2: Extensive glacier mass balance and flow studies within major climatic zones for improved process understanding and calibration of numerical models
Full parameterisation of coupled numerical energy/mass balance and flow models is based on detailed observations for improved process understanding, sensitivity experiments and extrapolation to areas with less comprehensive measurements. Ideally, sites should be located near the centre of the range of environmental conditions of the zone which they are representing. The actual locations will depend more on existing infrastructure and logistical feasibility than on strict spatial guidelines, but there is a need to include a broad range of climatic zones (such as tropical, subtropical, monsoon-type, mid-latitude maritime/continental, subpolar, polar).

Tier 3: Determination of regional glacier volume change within major mountain systems using cost-saving methodologies
There are numerous sites that reflect regional patterns of glacier mass change within major mountain systems, but they are not optimally distributed (Cogley and Adams 1998). Observations with a limited number of strategically selected index stakes (annual time resolution) combined with precision mapping at about decadal intervals (volume change of entire glaciers) for smaller ice bodies, or with laser altimetry/kinematic GPS (Ahrendt et al. 2002) for large glaciers constitute optimal possibilities for extending the information into remote areas of difficult access. Repeated mapping and altimetery alone provide important data at a lower time resolution (decades).

Tier 4: Long-term observations of glacier length change data within major mountain ranges for assessing the representativity of mass balance and volume change measurements
At this level, spatial representativity is the highest priority. Locations should be based on statistical considerations concerning climate characteristics, size effects and dynamics (e.g., calving, surge, debris cover). Long-term observations of glacier length change at a minimum of about ten sites within each of the important mountain ranges should be measured either in-situ or with remote sensing techniques at annual to multi-annual frequencies.

Tier 5: Glacier inventories repeated at time intervals of a few decades by using satellite remote sensing
Continuous upgrading of preliminary inventories and repetition of detailed inventories using aerial photography or – in most cases – satellite imagery should make it possible to attain global coverage and to serve as validation of climate models (Beniston et al. 1997). The use of digital terrain information in GIS greatly facilitates automated procedures of image analysis, data processing and modelling/interpretation of newly available information (Kääb et al. 2002, Paul et al. 2002, Paul 2004). Preparation of data products from satellite measurements must be based on a long-term program of data acquisition, archiving, product generation, and quality control, which is currently the purpose of the international Global Measurements from Space (GLIMS; www.glims.org) project (cf. Kieffer et al. 2000, Bishop et al. 2004, Raup et al. accepted).

The classification of the glacier information from the historical expansion of the Alpine monitoring network exemplifies the implementation of the system of tiers for the Euorpean Alps, see Figure 3.11. This integrated and multi-level strategy aims at combining in-situ observations with remotely sensed data, process understanding with global coverage and traditional measurements with new technologies. According to Haeberli (2004), Tiers 2 and 4 mainly represent traditional methodologies which remain fundamentally important for a deeper understanding of the processes involved, as training components in environment-related educational programmes, and as unique demonstration objects for a wide public, whereas Tiers 3 and 5 constitute wide-open doors for the application of new technologies (cf. Kääb 2005).

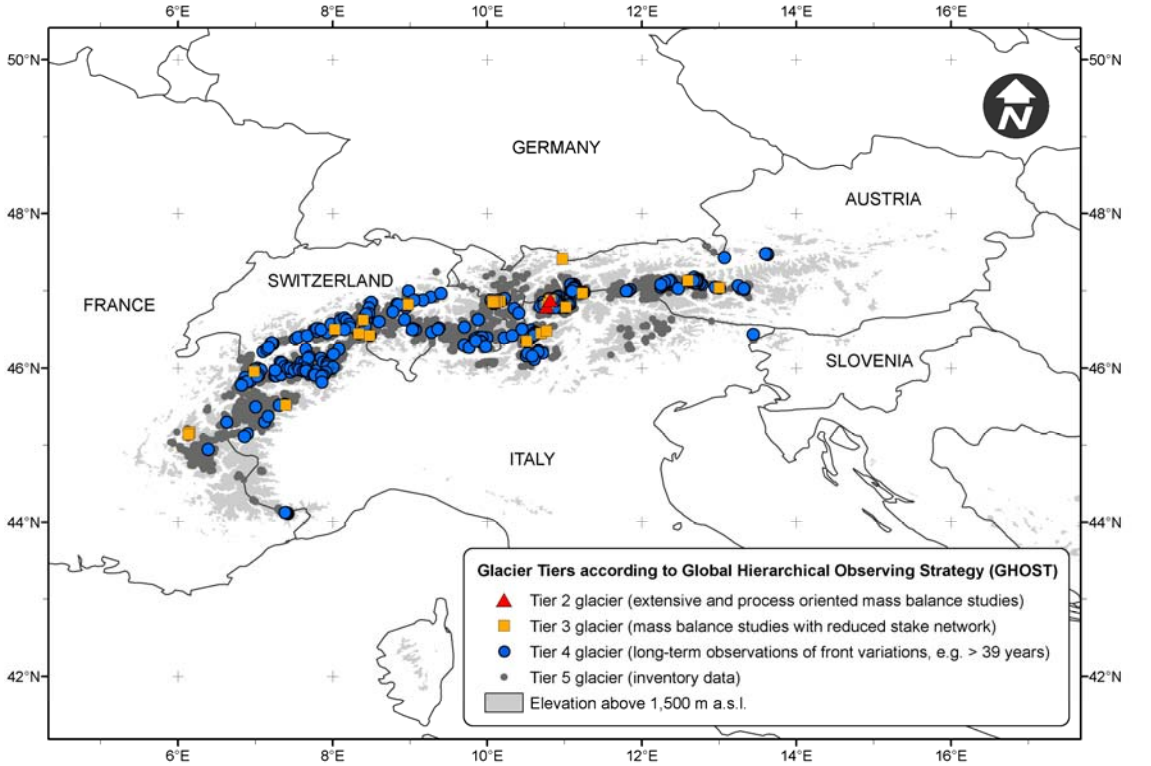

Figure 3.11: Historically-developed Alpine glacier monitoring network. The available glacier data is exemplarily classified according to the different Tiers of the GHOST. Explanation is given in the text. Background: European HYDRO1k-DEM, elevations above 1,500 m a.s.l. are shaded in grey. Source: LP DAAC, located at the USGS EROS, http://LPDAAC.usgs.gov. Country boundaries provided by ESRI.

3.4 Alpine glacier fluctuations before 1850

Direct measurements (glaciological method) of glacier volume, surface area and fluctuations in length are available for the past 120 years for the European Alps (Haeberli & Zumbühl 2003, Haeberli in press). Glacier fluctuations before that time can be reconstructed using various methods of differing precision and covering different periods back into the past. Thereby, the historical methods (past 500 to 700 years) involve the collection of data from written and pictorial historical records. The archaeological methods (past 700 to 800 years) analyse traces from human activity correlated with glacier development. The glacio-morphological methods (entire Holocene) entail the search for dateable organic material such as soils and crushed trees as well as the examination of the growth rings of fossil trees (dendrochronology) within the edge of glacier forefields (Holzhauser et al. 2005). Moraine sequences can be mapped and dated by absolute and relative age-determination methods, such as Schmidt-hammer rebound, weathering rind thickness, lichenometry, luminescence dating, cosmogenic (exposure) dating and radiocarbon dating.

At the LGM (about 21,000 y BP), glaciation covered an area of about 150,000 km^2 in the European Alps (Keller & Krayss 1998). Already in the early 19th century, shortly after the emergence of the ice age theory, rough estimations of this glaciation were made by Venetz (1821) and Agassiz (1841). Sketches and maps of the extent of the Alpine glaciation at the LGM can be found in Klebelsberg (1948), Habbe (1977; see Figure 3.12), Glückert (1987), Ehlers (1994), Gudmundsson (1994) and Keller & Krayss (1998). Regional maps were published in Jäckli (1962, 1970), Haeberli & Penz (1985), Pugin & Wildi (1995) and Keller & Krayss (1998) for Switzerland, De Beaulieu et al. (1988) for France and Van Husen (1978), Weinhart (1973) and Scholz (1995) for the eastern Alps. A detailed overview of literature about the glaciation of the LGM in the European Alps is given by Benz (2003).

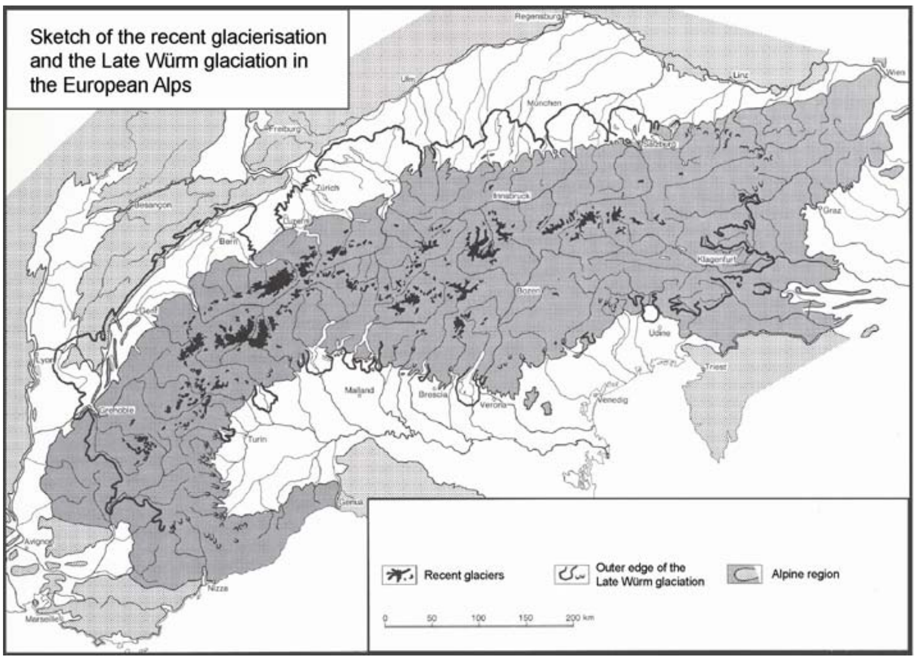

Figure 3.12: Extents of the recent glacierisation and of the glaciation at the LGM (about 21,000 y BP). During the Würm glaciation, glaciers advanced far into the lowlands in the north-western and northern parts of the Alps, whereas they stayed within the Alpine range in the east. Slightly modified after Habbe (1977) in Bachmann (1978).

With the increasing temperature after the LGM, glaciers started to retreat between 17,000 and 10,000 y BP (Keller 1988). Moraine sequences mark the chronology of intermittent re-advances during the general retreat from the forelands back into the Alpine valleys. The moraines of the YD 'Egesen Stadial' (about 11,000–10,200 y BP) are conspicuous features in many parts of the Alps and have been mapped extensively (e.g., Mayr & Heuberger 1968, Kerschner 1978, Furrer et al. 1987). They represent the last of the Late Glacial cold event periods prior to the many advances of Holocene age, which characteristically reached only the LIA extent (e.g., Patzelt and Bortenschlager 1973, Holzhauser 1984, Kerschner et al. 2000).

Hormes et al. (2001) presented a radiocarbon data set of wood and peat samples from six glaciers in the Central Swiss Alps, documenting eight Holocene phases (9,910–9,550, 9,010–7,980, 7,250–6,500, 6,170–5,950, 5,290–3,870, 3,640–3,360, 2,740–2,620 and 1,530–1,170 y BP) of glacier extents smaller than 1990. However, recent archaeological findings on crest/saddle sites like the Oetztal iceman at the Austrian-Italian border, the wooden bows at Lötschen Pass and the Neolithic leader cloths at Schnidejoch in the Swiss Alps show that the ice cover at theses sites have never been smaller during the last 5,000 years than today (Haeberli et al. 2004, Grosjean et al. 2006)). Holzhauser et al. (2005) reconstructed glacier fluctuations over the last 3,500 years (Figure 3.13) from a data set of tree-ring width, radiocarbon and archaeological data, as well as historical sources. They found nearly synchronous advances of Great Aletsch, Gorner and Lower Grindelwald glaciers in the Swiss Alps at 1000–600 BC and (AD) 500–600, 800–900, 1100–1200 and during the so-called Litte Ice Age (1300–1860). Similar studies, even though not extending that far back in time, do exist for instance for the Fiescher, Schwarzberg, Allalin, Fee, Rosenlaui and Upper Grindelwald glaciers in Switzerland (Holzhauser & Zumbühl 2003), Hintereisferner (Nicolussi 1994), Pasterze and Gepatschferner (Nicolussi & Patzelt 2000) in Austria, and for the Forni (Pelfini 1988) and Madaccio (Pelfini 1999) glaciers in Italy.

During the six centuries of the LIA, the climate was generally favourable for glaciers (Maisch 1992), but, as Pfister (2005) emphasised, it was not continuously cold (see also Casty et al. 2005). The cold phases, interrupted repeatedly by phases of 'average climate', resulted in three main glacier advances in the 14th century, in the 17th century and the last one culminating around 1850. Most Alpine glaciers reached their Holocene maximum extent during this last advance and destroyed earlier moraines (Maisch 1992, Holzhauser & Zumbühl 2003). However, in some regions the moraines from 1820 mark this maximum extent (e.g., Wetter 1987, Holzhauser & Zumbühl 2003). Digital inventories of the 1850 glacier extent are available for Switzerland (Maisch et al. 2000) and Austria (partly published in Gross 1983, 1987). In Italy and France this information is available for individual regions (e.g., Orombelli & Mason 1997, Pelfini 1999, Cossart et al. in press).

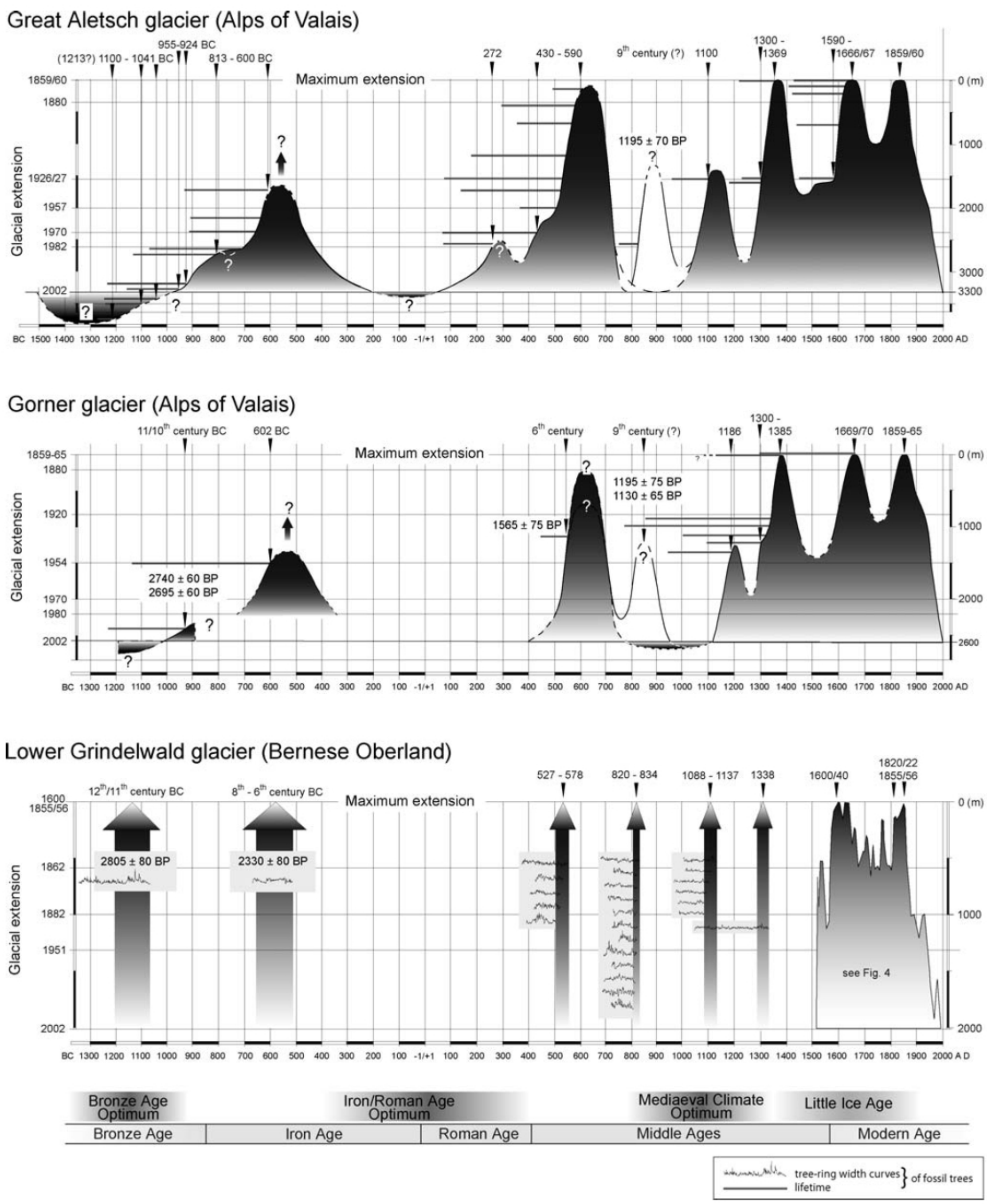

Figure 3.13: Fluctuations of the Great Aletsch, the Gorner and the Lower Grindelwald glaciers over the last 3,500 years. Figure from Holzhauser et al. (2005).

4 Summary of research

The work of this thesis consists primarily of the revision and extension of the existing Alpine glacier data set available from WGMS (*Paper I*), the spatio-temporal analysis of this new, extensive data set (*Papers I*, *II*, *III*), the comparison of Alpine glacier fluctuations with glacier changes in other mountain ranges (*Papers II* and *III*), the development of a numerical model to reproduce present glacierisation from gridded climatologies (*Paper IV*), as well as the simulation of past (*Paper IV*) and future (*Papers IV* and *V*) glacierisations of the entire European Alps. The results of this research and their discussion are published in international scientific journals (*Papers II*, *III, IV* and *V*), or as a peer-reviewed article in a scientific book (*Paper I*). Here in *Chapter 4*, a brief summary of each paper is given. A full version of the papers can be found in *Part B*.

4.1 Paper I

Zemp, M., Paul, F., Hoelzle, M. & Haeberli, W. (in press): *Glacier fluctuations in the European Alps 1850–2000: an overview and spatio-temporal analysis of available data.* In: Orlove, B., Wiegandt, E. & Luckman, B. (eds.): The darkening peaks: Glacial retreat in scientific and social context. University of California Press.

This paper briefly summarises the history of internationally coordinated glacier observations and corresponding publications, gives an overview of the revised and extended Alpine glacier data set back to 1850, presents the result of a spatio-temporal analysis of this data and discusses the representativity of front variation and mass balance series for the entire Alpine glacierisation.

In the European Alps, the growth of the glacier monitoring network over time has resulted in an unprecedented glacier data set with excellent spatial and temporal coverage. The WGMS compiled information on spatial glacier distribution from approximately 5,150 Alpine glaciers and fluctuation series from more than 670 of these glaciers (i.e., more than 25,350 annual front variation and 575 annual mass balance observations at the time of this analysis) dating back to 1850. National inventories are able to provide complete Alpine coverage for the 1970s, when the glaciers covered a total area of 2,909 km^2. This inventory, together with digital outlines from the Swiss Glacier Inventory, is used to extrapolate total Alpine glacier-covered areas in 1850 and 2000 which amount to about 4,470 km^2 and 2,270 km^2, respectively. This corresponds to an overall loss from 1850 to the 1970s of 35%, and almost 50% by 2000. Annual mass balance and front variation series provide a better time resolution of glacier fluctuations over the past 150 years than the inventories. During the general retreat, intermittent periods of glacier re-advances in the 1890s, 1920s and 1970–1980s can still be seen. Increasing mass loss, rapidly shrinking glaciers, disintegrating and spectacular tongue retreats are clear signs of the atmospheric warming observed in the Alps during the last 150 years and the acceleration observed over the past two decades. However, in the short term or at a regional scale, glaciers show a highly individual variability. Glacier behaviour depends not only on regional climate but also on local topographic effects which complicate the extraction of the climate signal from glacier fluctuations.

While inventory data contain information on spatial glacier distribution at certain times, fluctuation series provide temporal information at specific locations. Continuity and representativity of fluctuation series are thus most relevant for the planning of glacier monitoring. Furthermore, modelling should be enhanced and integrated into monitoring strategies. It is of major importance to continue with long-term fluctuation measurements and to extend the series back in time with reconstructions of former glacier geometries. Additionally, it is necessary to integrate glacier modelling as well as reconstruction activities into the global framework of the GLIMS project and the WGMS.

4.2 Paper II

Hoelzle, M., Chinn, T., Stumm, D., Paul, F., **Zemp, M.** & Haeberli, W. (accepted): *The application of glacier inventory data for estimating past climate-change effects on mountain glaciers: a comparison between the European Alps and the Southern Alps of New Zealand*. Global and Planetary Change, special issue on climate change impacts on glaciers and permafrost.

In this study, glacierisations of the mid-1970s in the European Alps and in the Southern Alps of New Zealand are compared. A simple parameterisation scheme from Haeberli & Hoelzle (1995) is applied to the two inventories. This approach considers a step change in climate and following, a full dynamic glacier response until new steady-state conditions have been achieved. It allows the estimation and comparison of glaciological parameters (e.g., mass balance at glacier tongue, ice thickness, mean specific mass balance, volume, reaction and response time) from basic inventory data, such as glacier area, length, minimum and maximum elevations.

In the European Alps and the Southern Alps of New Zealand, there are for the mid-1970s a total of about 5,150 and 3,130 perennial surface ice bodies, covering 2,909 km^2 and 1,139 km^2, respectively. The parameterisation scheme is applied to glaciers larger than 0.2 km^2, which are 1,763 (35%) for the European Alps and 702 (22%) for the Southern Alps of New Zealand, covering 88% and 86% of the corresponding total surface areas. Multiplication with average glacier thickness yielded total volumes of 126 km^3 and 67 km^3, respectively. The calculated area change between ‘1850’ extents and the mid-1970s are -35% for the European Alps and -49% for the New Zealand Alps, with a corresponding volume loss of 48% and 61%, respectively. From data on measured cumulative length change, an average mass balance for the investigated period could be determined at -0.33 m w.e. per year for the European Alps, and -1.25 m w.e. for the ‘wet’ and -0.54 m w.e. per year for the ‘dry’ glaciers of the New Zealand Alps. However, there is some uncertainty in several factors, such as the values used in the parameterisation scheme of mass balance gradients, which, in New Zealand vary between 5 and 25 mm m^{-1} (about 7.5 mm m^{-1} in the European Alps). Glacier behaviour after the mid-1970s was sometimes quite different between the two mountain ranges. After a period of glacier advance in the 1970s and early 1980s glaciers in the European Alps experienced a strong if not accelerated retreat until present day. In contrast, all glaciers in New Zealand have experienced an overall mass gain, and some advanced strongly during the 1990s, coming to a halt at the beginning of the new century. The glaciers in the New Zealand Alps will probably react more sensitively to a future temperature increase than those in the European Alps. Not only because of their greater climate sensitivity, as expressed by the mass balance gradient, but also because of the generally low altitude of the ELA in the New Zealand Alps, promoting a higher percentage of rain rather than snowfall in the future.

The potential of this parameterisation of existing inventories lies especially in its ability to quantitatively infer past average decadal to secular mean specific mass balances for unmeasured glaciers. However, drastic changes in geometry and downwasting rather than active retreat might be the dominant processes of glacier reaction to predicted fast climate change. This implies that parameterisation schemes like the presented one would no longer be applicable and new approaches must be developed and applied.

4.3 Paper III

Zemp, M., Frauenfelder, R., Haeberli, W. & Hoelzle, M. (2005): *Worldwide glacier mass balance measurements: general trends and first results of the extraordinary year 2003 in Central Europe*. Data of Glaciological Studies [Materialy glyatsiologicheskih issledovanii], 99: p. 3–12.

International monitoring strategies aim at combining in-situ measurements with remotely sensed data, process understanding with global coverage, and traditional measurements with new technologies by using an integrated and multi-level approach. Mass balance observations are an important component in these strategies, as they are a direct and undelayed signal of annual atmospheric conditions. In the present study, worldwide continuous mass balance records for the period 1980–2001 are analysed. Observed rates of change and corresponding trends from nine Alpine glaciers are compared with the ones from the other eight mountain ranges. In addition, the extraordinary warm year 2003 in Central Europe and its impact on the Alpine glacierisation are investigated.

Mean (annual) specific net balance for the nine mountain ranges (Cascades, Alaska, Andes, Svalbard, Scandinavia, Alps, Altai, Caucasus and Tien Shan) during the period 1990–1999 amounts to -482 mm w.e., a value three times higher than the corresponding value for the previous decade from 1980–1989 (-149 mm w.e.). For the time period 1980–2001, the mean specific net balance averaged roughly -0.3 m w.e. per year, resulting in a total thickness reduction of approximately 7 m w.e. since 1980. The corresponding values (1980–2001) for the Alps were about -0.6 m w.e. per year and a total ice loss of almost 13 m w.e. In the European Alps, the beginning of the hydrological year 2003 was characterised by high accumulation lasting until the end of February. After that, the Alpine weather was dominated by anticyclonic weather types, which are characteristic of high temperatures, high short-wave radiation, few clouds and low precipitation. This reversed the above-average accumulation of the first half of the winter into low snow depths in late winter, fast melting in spring and an unusually early start of the ablation season. The ablation period lasted for more than three months, again dominated by anticyclonic influences. The summer heatwave with its centre over France led to a summer 3-month (JJA) temperature anomaly of 5.1 °C, corresponding to an offset of 5.4 standard deviations. These weather conditions resulted in a mean specific net balance of Alpine glaciers of -2.5 m w.e. in 2003. This is more than 50% more than the value during the previous record year of 1998 (-1.6 m w.e.) and roughly four times greater than the Alpine average loss of 0.6 m w.e. per year since 1980. In addition, this value is almost one order of magnitude higher than the global mean annual net balance of -0.3 m w.e. recorded during the already exceptionally warm period 1980–2001. Assuming a total Alpine glacierized area of 2,909 km^2 in the mid-1970s, relative loss in glacier area from mid-1970s to 2000 of 22%, an ice loss of about 2.5 m w.e. in 2003 and a total ice volume of approximately 75 km^3 before 2003, estimated total glacier volume loss in 2003 corresponds to roughly 5–10% of the ice volume still present before that year.

In comparison with possible future European climate scenarios, climatic conditions (temperature and precipitation) in summer 2003 were not unlike those simulated by the climate scenario runs for the period 2071–2100. The question whether the extraordinary summer of 2003 was only an aberration of today’s climate, or already a harbinger of things to come, is hard to answer. However, the summer of 2003 and its impact on Alpine glaciers may well serve as a case study for such a possible future climate.

4.4 Paper IV

Zemp, M., Hoelzle, M. & Haeberli, W. (accepted): *Distributed modelling of the regional climatic equilibrium line altitude of glaciers in the European Alps*. Global and Planetary Change, special issue on climate change impacts on glaciers and permafrost.

The equilibrium line altitude of a glacier is a theoretical line which defines the altitude at which annual accumulation equals ablation. It represents the lowest boundary of the climatic glacierisation and, therefore, is an excellent indicator of climate variability. In this work, a simple approach that is able to model the glacier distribution at high spatial resolution (about 100 m) over entire mountain ranges is presented, the sensitivity of the Alpine glacierisation to changes in temperature and precipitation is investigated, and the possible impact of a climate change scenario on Alpine glacierisation in the decades to come is demonstrated. To achieve this, a simple approach for modelling the glacier distribution at high spatial resolution is introduced, using a minimum of input data. An empirical relationship between precipitation and temperature at the ELA_0, is derived from direct glaciological mass balance measurements. Using geographical information systems (GIS) and a digital elevation model (DEM), this relationship is then applied over a spatial domain, to a so-called distributed modelling of the regional climatic ELA_0 ($rcELA_0$) and the corresponding climatic accumulation area (cAA) for 1971–1990 over the entire European Alps.

The sensitivity study shows that for Alpine glaciers, a change in 6-month summer temperature by ±1 °C would be compensated by an annual precipitation increase/decrease of about 25%. The modelled cAA of 1971–1990 agrees well with accumulation areas of mapped glacier outlines from the 1973 Swiss Glacier Inventory and covers an area of 3,059 km^2 over the entire Alps. As the cAA is simply the terrain above the $rcELA_0$, it does not distinguish between glacier surface and ice-free rock walls. In a first order approach, this can be corrected by applying of slope-dependent function of glacier fraction to the modelled cAA. The thus corrected cAA equals 1,950 km^2 and corresponds to an AAR_0 of 0.67 of the measured total Alpine glacier area in the 1970s, which was 2,909 km^2. Assuming a warming of 0.6 °C between 1850 and 1971–1990 leads to a mean $rcELA_0$ rise of 75 m and a corresponding cAA reduction of 26%. This latter value is somewhat lower than the loss in glacier area (35%) as estimated from glacier inventory data from 1850 and the 1970s, and suggests that either the model is not perfectly able to reproduce the area loss between 1850 and the 1970s, or that a rise in temperature by 0.6 °C cannot completely explain the corresponding glacier shrinkage. A further rise in temperature of 3 °C accompanied by an increase in precipitation of 10% leads to a further mean rise of the $rcELA_0$ of about 340 m and reduces the cAA of 1971–1990 by about 75%. As a consequence, many regions become ice-free and remaining accumulation areas of glaciers are strongly reduced and disintegrated.

The presented approach is an excellent complement to distributed mass balance models. As the distributed $rcELA_0$-model requires only a minimum amount of input data to compute the $rcELA_0$ over the entire Alps, distributed mass balance models can then be used to account for further important components of the energy balance (e.g., solar radiation, albedo, turbulent fluxes, mass balance-altitude feedback) and local, topographic effects (e.g., shading, avalanches, snowdrift) within individual catchments. Furthermore, distributed modelling of the $rcELA_0$ can potentially contribute to the current efforts to include glacier altitude-area distribution of past, present and future glacier states in regional climate models.

4.5 Paper V

Zemp, M., Haeberli, W., Hoelzle, M. & Paul, F. (submitted): *Alpine glaciers to disappear within decades?* Geophysical Research Letter.

In this study, an integrated approach, combining in-situ measurements, remote sensing techniques and numerical modelling is applied to the European Alps. These techniques allow past as well as potential evolutions of a glacier ensemble within an entire mountain chain to be assessed quantitatively. Glacier fluctuations from 1850 to 2000 are analysed from earlier and recent inventories, together with compilations on past glacier fluctuations. Potential future area changes for the entire Alps are estimated using two independent methods. The first method is a purely empirical one that relates documented rates of area change for altitudinal bands to scenarios of glacier shrinking. The second approach (based on *Paper IV*) is a statistically calibrated and distributed model of glacier ELA that utilises an empirical relation between 6-month summer air temperature and annual precipitation at the ELA_0. With this model, the cAA is computed over the entire Alps for the reference period (1971–1990) and for the entire range of temperature and precipitation scenarios for the 21st century, as published by IPCC (2001). Finally, past, present and future ice volumes are calculated by multiplying reconstructed/measured mass balance values with the average surface area of a given time period.

Alpine glaciers lost almost 50% in area from 1850 to 2000. The area reduction between 1975 and 2000 is about 22%, mainly occurring after 1985. Disintegration and 'downwasting' have been predominant processes of glacier decline during the most recent past. The scenarios of future area losses illustrate that the scenario of 'accelerated loss' (area reduction for the period 1975–2000) would drastically reduce Alpine glacier areas within this century and that the scenario of extreme ice loss (using doubled 1985–2000 loss rates) would cause most of the presently existing glaciers in the Alps to disappear within decades as large parts of the ice is located below 3,000 m a.s.l. The impact on glacier areas as related to scenarios of temperature and precipitation show that a 3 °C warming of summer air temperature would reduce the currently existing Alpine glacier cover by some 80%, or up to 10% of the glacier extent of 1850. In the event of a 5 °C temperature increase, the Alps would become almost completely ice-free. Annual precipitation changes of ±20% would modify such estimated percentages of remaining ice by a factor of less than two. Volume loss between 1975 and 2000 is calculated from cumulative mass balance of -12 m w.e. over a mean Alpine glacier area of 2590 km^2 to be 30 km^3. Total Alpine ice volumes can be estimated roughly as 200 km^3 in 1850, 105 km^3 in 1975 and 75 km^3 in 2000. The average mass balance of -2.5 m w.e. in the extreme year 2003 eliminated an estimated 8% of the remaining Alpine ice volume within one single year. Such extremely hot and dry summers not only induce strong positive feedbacks, but also eliminate increasing percentages. It is likely that five rather than ten repetitions within the coming decades of conditions as in 2003 would bring about the scenario of widely deglaciated Alps.

By simply looking at the evolution of glaciers in the mountain ranges of the world, coming generations will be able to define and to physically see which scenario of climate change has taken place.

5 General discussion

In this chapter a discussion on the main tasks and findings of the thesis is given. A more detailed discussion of methods and results can be found in the corresponding sections of the individual papers, appended in *Part B* of this work.

5.1 Revision and extension of the Alpine glacier data set

The starting points of this work were the *World Glacier Inventory* and the *Fluctuations of Glaciers* databases available at WGMS. The data model and corresponding attributes are described in IAHS(ICSI)/UNEP/UNESCO (1989) and Hoelzle & Trindler (1998). In a first step, the existing data model was revised and extended to fulfil the requirements of relational (geo-) database management systems (cf. Jones 1997). Conceptual and internal/logical data models were designed for Oracle and MS Access databases and implemented in MS Access. The revised data model allows direct connection of a GIS (ArcGIS 9.x from the GIS vendor ESRI) to the database, which facilitates visualisation, control and analysis of the data by GIS techniques. The integration of the extended data set into the database of the WGMS ensures public availability of the data and continuity of the new data far beyond the time range of this thesis and the ALP-IMP project.

In a second step, the existing data set was checked and extended by data found in the literature and by the integration of digital data collections of other Alpine glaciologists, in collaboration with national correspondents and principal investigators of the WGMS. Besides the inventory data on the approximately 5,150 Alpine glaciers (i.e., area, length and altitude range), the new data set contains (at the end of the project) fluctuation series from more than 920 Alpine glaciers dating back to 1850, including more than 25,350 annual front variation observations (from 677 glaciers) and 690 annual mass balance measurements (from 25 glaciers). Compared to the old data set (before 2003), this corresponds to an increase in available data by 65% for the number of glaciers with fluctuation data, by 34% for front variation observations and by 32% for annual mass balance measurements.

The third step consisted in the control and correction of the available data. In this process, visualisation of the data and plausibility checks facilitated the detection of errors. However, the subsequent adjustments were laborious, time-consuming and often only possible in close collaboration with the responsible institutes or local investigators of the corresponding glacier. Figure 5.1 exemplifies control and correction of latitude and longitude of glacier label points. Glacier coordinates were visualised in a GIS and coloured according to their attribute ‘political unit’. In comparison with digital country outlines and satellite images, errors could be detected easily and were then corrected with the help of topographic maps and information from the World Glacier Inventory (IAHS(ICSI)/UNEP/UNESCO 1989).

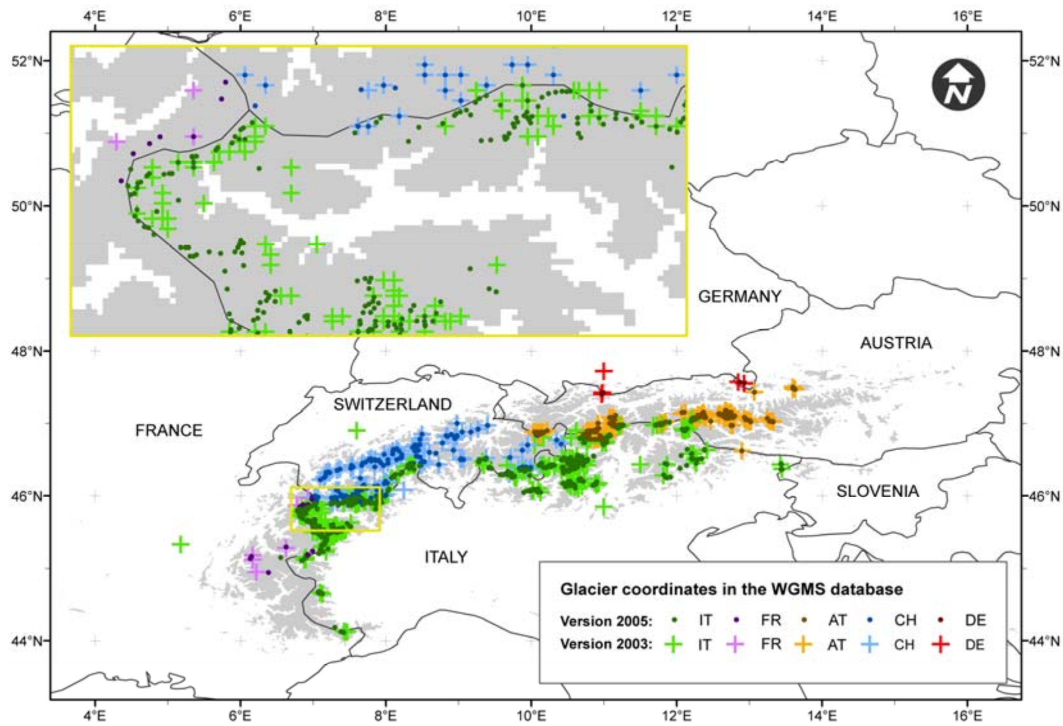

Figure 5.1: GIS-based control and correction of glacier coordinates. Original (crosses) and corrected (dots) glacier label points are coloured according to the glacier's political unit. The coordinates from more than 28% of the 923 glaciers with fluctuation measurements were corrected during this project.

5.2 Alpine glacier fluctuations after 1850

The inventory of the 1970s and the extrapolated area estimates for 1850 and 2000 show the dramatic dimension of glacier shrinkage in the Alps. Despite the high degree of variability of the individual glaciers, the European Alps have experienced a 50% decrease in ice coverage over the last 150 years. The area loss per decade (in percent) between the 1970s and 2000 is almost three times faster than the loss of ice between 1850 and the 1970s. This might be due to the continued or even enhanced climate forcing and to the different length of the time periods considered. However, variations in glacier frontal position provide a higher time resolution of the glacier retreat over the past 150 years. Though glaciers have generally been retreating since 1850, there have been several periods of documented re-advances in the 1890s, 1920s and 1970–1980s (Patzelt 1985, Müller 1988, Pelfini & Smiraglia 1988, Reynaud 1988 and Haeberli et al. 1989b). The area reduction after the 1970s mainly occurred after 1985 (see also, Paul et al. 2004). Thus, the acceleration of glacier retreat in the last two decades was even more pronounced. Mass balance measurements are available only for the last five decades and confirm the general trend of glacier shrinkage. While some glaciers gained mass between 1960 and 1980, ice loss has accelerated over the last two decades. The mean specific (annual) net balance of the 1980s is 18% more negative than the average of 1967–2001, and the value of the 1990s even doubles the average ice loss of 1967–2001. Most recent mass balance data show a continuation of the acceleration trend after 2000 with a peak in the extraordinary year of 2003 when the ice loss of the nine Alpine reference glaciers was about 2.5 m w.e. – exceeding the average of 1967–2001 by a factor of close to seven. Estimated total glacier volume loss in the Alps in 2003 corresponds to 5–10% of the remaining ice volume before that year. The acceleration of the glacier shrinkage after 1985 indicates transition towards a rapid downwasting rather than a dynamic glacier response to a changed climate (cf. Paul et al. 2004). Figure 5.2 shows ELA fluctuations from mass balance measurements relative to the glacier elevation range. The rise of the average ELA after 1980 documents the imbalance of glaciers from steady-state conditions. Glaciers with repeated ELA-values above the maximum elevation will not be able to remain under the current conditions.

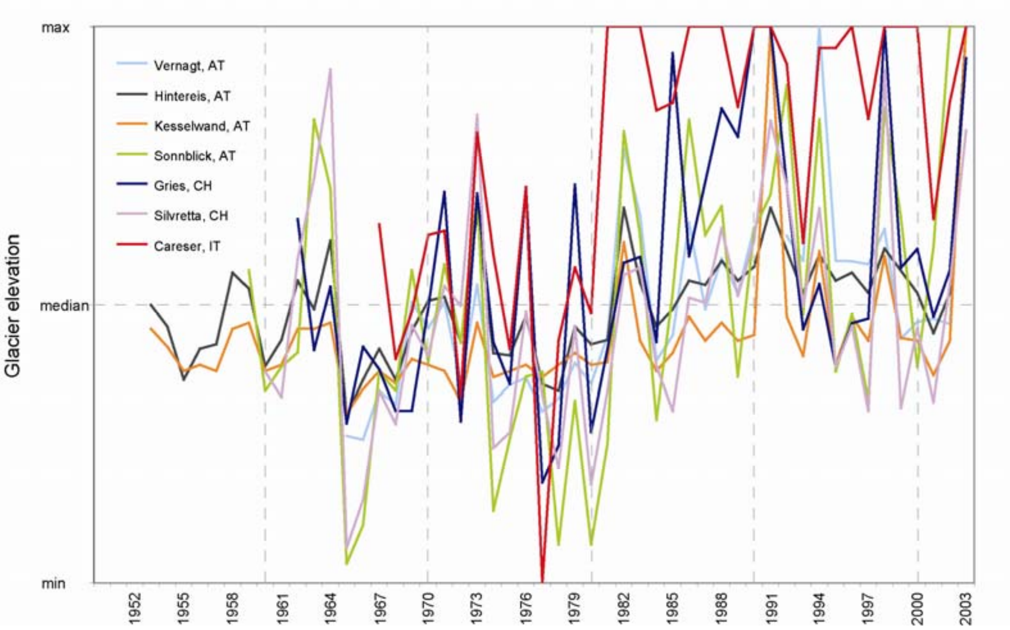

Figure 5.2: Alpine ELA fluctuations from 1953 to 2003. Annual ELA fluctuations from direct mass balance measurements are plotted relative to the glacier elevation range of 1969/1973/1980 for Austrian/Swiss/Italian glaciers. The median elevation corresponds to the altitude that divides the glacier surface area into two equal parts. No ELA values are available for French glaciers.

The general glacier retreat since 1850 corresponds well with the observed warming trend in this period (see also Oerlemans 1994, Maisch et al. 2000, Oerlemans 2001 p. 110–111). However, the onset of the Alpine glacier retreat after 1850 might have been triggered by a negative winter precipitation anomaly (relating to the mean of 1901–2000) during the second half of the 19th century (Wanner et al. 2005). The intermittent periods of glacier advances in the 1890s, 1920s and 1970–1980s can be explained by earlier wetter and cooler periods, with reduced sunshine duration and increased winter precipitation (Patzelt 1987, Schöner et al. 2000, Laternser & Schneebeli 2003). Schöner et al. (2000) concluded from the investigation of a homogenised climate data set and from measured mass balance data from the Austrian part of the eastern Alps, that the more positive mass balance periods show a high correlation with winter accumulation and a lower correlation with summer temperature, while more negative mass balance periods correlate extremely well with summer temperature and show no correlation to winter accumulation. In addition, they find that the positive mass balance period between 1960 and 1980 was characterised by negative winter North Atlantic Oscillation index values, which caused an increase in the meridional circulation mode and a more intense north-westerly to northerly precipitation regime (cf. Wanner et al. 2005). Another influence might also come from the dimming (i.e., the decline of solar radiation) reported for the period of about 1960–1990 and the subsequent brightening during the 1990s (cf. Wild 2005). The observed trend of increasingly negative mass balances since 1980 is consistent with accelerated global warming and correspondingly enhanced energy flux towards the earth’s surface (IUGG(CCS)/UNEP/UNESCO 2005).

5.3 Comparison of glacier changes in the European Alps with other mountain ranges

The Southern Alps of New Zealand are particularly interesting for a comparison with the European Alps, because they are situated at similar latitude (however, having quite different climatic conditions) and also have a highly accurate glacier inventory for the 1970s. A parameterisation scheme was applied to these inventories to estimate glacier changes between the '1850 extent' and the 1970s. The calculated area change is -49% for the New Zealand Alps and -35% for the European Alps, with a corresponding estimated volume loss of -61% and -48%, respectively. From cumulative measured front variation data an average mass balance for the investigated period was assessed at -0.33 m w.e. per year for the European Alps and -1.25/-0.54 m w.e. per year for glaciers in the 'wet'/'dry' regions of the Southern Alps of New Zealand. Glacier behaviour after the mid-1970s was sometimes quite different between the two mountain ranges. After a period of glacier re-advances in the 1970s and early 1980s in the European Alps, the glaciers experienced a strong if not accelerated retreat until present. In contrast, all glaciers in New Zealand have experienced an overall mass gain, and some advanced strongly during the 1990s. This growth trend came to an end at the beginning of the new century.

Hoelzle et al. (2003) compared mean specific mass balances estimated from long-term measurements of cumulative glacier length change for different mountain regions, ranging from dry, continental-type climate conditions (e.g. Altai) over transitional climates (e.g. Caucasus) to humid, maritime-type conditions (e.g. Western Norway). They found that on average of the worldwide sample, mean specific mass balance since 1900 centres around an average value of about -0.25 m w.e. per year, and that the reconstructed rates of secular mass losses strongly differ between humid, maritime-type glaciers (> -0.5 m w.e. per year) and dry, continental-type glaciers (< -0.1 m w.e. per year). Thus, the sensitivity with respect to secular trends in global warming of maritime-type glaciers is much higher than that of continental-type glaciers (cf. also Kuhn 1981, Oerlemans 2001), with the European Alps somewhere in between.

Since 1980 continuous mass balance measurements have been available for 30 glaciers in nine mountain ranges (Cascades, Alaska, Andes, Svalbard, Scandinavia, Alps, Altai, Caucasus, Tien Shan). For the time period from 1980–2001, the mean specific net balance of these mountain regions averaged roughly -0.3 m w.e. per year, resulting in a total thickness reduction of approximately 7 m w.e. since 1980. The corresponding values for the European Alps were about -0.6 m w.e. per year and a total ice loss of almost 13 m w.e. In addition to the Cascade Mountains with a total ice loss of more than 19 m w.e. over this period, the Alpine reference glaciers show the greatest cumulative mass loss after 1980.

The European Alps represent a low-latitude mountain chain with less dense glacier cover than elsewhere. Numerous medium-sized and quite steep mountain glaciers constitute the most important part of the area distribution frequency (Figure 5.3, left). Such small/steep glaciers adjust relatively quickly to changing climatic conditions and can, therefore, quite realistically be assessed assuming step-wise changes between equilibrium conditions (as for instance applied in Hoelzle et al. 2003, or in *Paper II*). However, such concepts are not applicable in every case. In heavily glacier-covered regions like Svalbard (Norway), Patagonia (Argentina, Chile) or the St. Elias Mountains (Alaska), relatively few large and rather flat valley glaciers dominate the frequency diagram of area distribution (Figure 5.3 right). Because long, flat valley glaciers have dynamic response times beyond the century scale (Haeberli & Hoelzle 1995, Jóhannesson et al. 1989), fast climate change primarily causes (vertical) thinning of ice rather than (horizontal) retreat and area reduction. For such cases, conditions far beyond equilibrium stages, perhaps even run-away effects from positive feedbacks (mass balance/altitude), must be envisaged (Haeberli 2004, Raymond et al. 2005): 'downwasting' or even 'collapse' of large ice bodies could become the most likely future scenarios related to accelerating atmospheric temperature rise in these areas, and have already been documented (e.g., Arendt et al. 2002, Rignot et al. 2003).

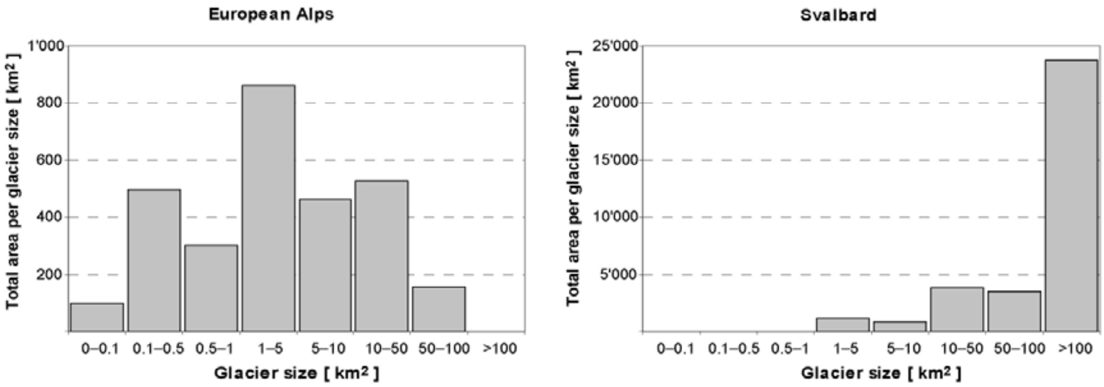

Figure 5.3: Glacier area distribution frequency in the European Alps and in Svalbard (Norway) in the 1970s. For Svalbard, glaciers with an area of less than 1 km^2 have not been considered in the inventory.

5.4 Modelling of past, present and future Alpine glacierisations

The model presented in *Paper IV* is a simple but robust approach that is able to compute the glacier distribution at high spatial resolution (about 100 m) over entire mountain ranges. It is based on an empirical relation between precipitation and temperature at the glacier ELA_0 derived from direct glaciological mass balance measurements, and enables the computation of the $rcELA_0$ and the corresponding cAA from gridded climatologies and a DEM. The model computes glacierisation from the intersection of the lowest boundary of the climatic glacierisation with the terrain, and does not depend on any (assumed) glacier AAR_0. The model can be used to calculate past, present or future glacierisation from climate data or it can be used in an inverse sense, to assess one of the climate parameters (temperature or precipitation) from known ELA_0 and the other climate factor. In contrast to the parameterisation scheme used in *Paper II*, the model of distributed $rcELA_0$ is independent of glacier dynamics (i.e., ice flow and corresponding variations in the front position), and, hence, is also applicable during continued or even accelerated climate forcing when glaciers react by downwasting rather than by dynamic retreat. However, for the assessment of total glacier area from the modelled cAA (or its change between two different glacier states), an AAR_0 (and its constancy over time) has to be assumed. The model does not take into account topographic effects, such as for instance snowdrift (cf. Purves et al. 1999), avalanches (cf. Gruber subm.) or solar radiation (cf. Dozier & Frew 1991), which lead to differences between the modelled $rcELA_0$ and the local topographic ELA_0 on glaciers. The influence of the uncertainty in $rcELA_0$ on the cAA is indirectly proportional to the slope of the terrain, i.e., the steeper the terrain, the smaller the corresponding error of the cAA. As the distributed $rcELA_0$ model requires only a minimum amount of input data to compute glacierisation over an entire mountain range, distributed mass balance models (e.g., Klok & Oerlemans 2002, Hock & Holmgren 2005, Paul et al. in press, Machguth et al. submitted) can then be used to account for further important components of the energy balance (e.g., solar radiation, albedo, turbulent fluxes, mass balance-altitude feedback) and local, topographic effects (e.g., shading, avalanches, snowdrift) within individual catchments with available data.

The model experiments (*Papers IV* and *V*) show that a change in 6-month summer temperature by ±1 °C would be compensated by a precipitation increase/decrease of 25%. A rise in 6-month summer temperature of 3 °C accompanied by an increase in annual precipitation of 10% would lead to a mean rise of the $rcELA_0$ of about 340 m and would reduce the cAA of 1971–1990 by about 75% (Figure 5.4). In the event of a 5 °C summer temperature increase, the Alps would become almost completely ice-free. Annual precipiation changes of ±20% would modify such estimated percentages of remaining ice by a factor of less than two. These results are in good agreement with studies based on different methods (Maisch et al. 2000, Haeberli & Hoelzle 1995) and shows that the possibility of Alpine glaciers disappearing by the end of the 21st century is far from slight.

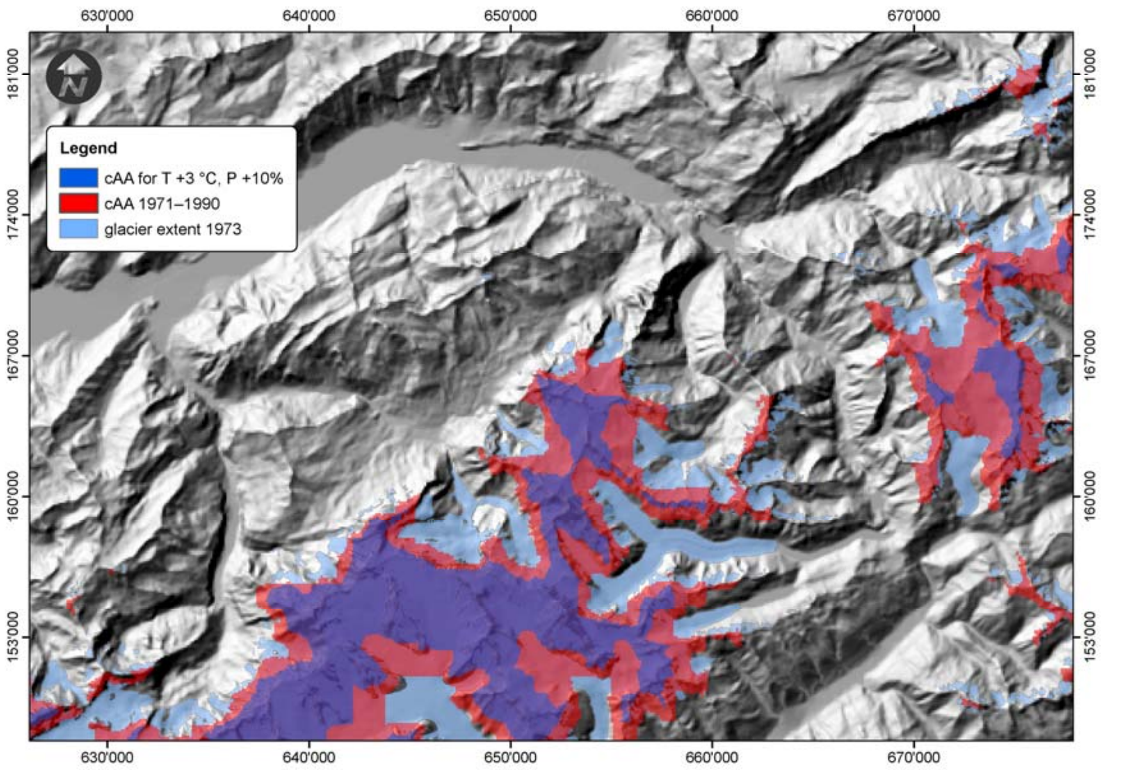

Figure 5.4: Future glacier scenario for the upper drainage basins of the Lütschine and Aare Rivers, Switzerland. A rise in 6-month summer temperature of 3 °C accompanied by an increase in annual precipitation of 10% would cause the $rcELA_0$ to rise by more than 300 m and strongly reduce the cAA of the reference model run (1971–1990). Background: Hillshading from DEM25 (reproduced by permission of swisstopo, BA067729). Glacier outlines from the Swiss Glacier Inventory (Kääb et al. 2002, Paul et al. 2002, Paul 2004).

5.5 Glacier monitoring in the European Alps

Glacier studies have a long tradition in the Alps that began with the establishment of systematic observation networks in the 1890s (Haeberli & Zumbühl 2003, Haeberli in press). In comparison to the rest of the world, the European Alps have the densest and most complete spatial glacier inventory through time (IAHS(ICSI)/UNEP/UNESCO 1989). Interestingly, the 1970s is the only period for which an Alpine inventory with total spatial coverage can be compiled, when most glaciers were relatively close to steady-state conditions (see also Patzelt 1985). The reconstructed glacier extents at the end of the LIA (around 1850) and the glacier outlines derived from multi-spectral satellite data around 2000 from the Swiss Glacier Inventory (Kääb et al. 2002, Paul et al. 2002, Paul 2004) cover the major parts and the full range of area classes of Swiss glacierisation. Thus they can be used to extrapolate Alpine glacierisation in 1850 and 2000, based on the assumption that the relative losses of the different area classes in Switzerland are representative of other Alpine countries. This, of course, is not necessarily the case. In addition, identification of individual glaciers throughout the different inventories (Müller et al. 1976, Maisch et al. 2000, Kääb et al. 2002 and Paul et al. 2002) is difficult due to different identification codes used and glacier retreat, which has caused severe changes in glacier geometry (e.g., tongue separation, disintegration). Together with the systematic differences between historical inventories (glacier coordinates and tabular attribute information) and modern satellite inventories (glacier outlines and tabular attribute information), the identification problem is a laborious but essential task that is still to be tackled.

The front variation series have the longest tradition in the observation network and, hence, are numerous and well distributed over the Alps, with a minimum number of observations in the south-western part of the Alps. Mass balance measurements started in the late 1940s. These measurements are labour-intensive and thus only available from 25 glaciers, with continuous series from nine glaciers since 1967. For the fluctuation series, length and completeness of the time series as well as representativity for the entire glacierisation are most relevant.

While inventory data contain information on spatial glacier distribution at certain times, fluctuation series provide temporal information at specific locations. Thus, glacier mass balance is the direct and undelayed signal of annual atmospheric conditions, whereas changes in length are an indirect, delayed, filtered but also enhanced signal (Haeberli 1998). Glacier behaviour depends not only on regional climate but also on local topographic effects which complicate the extraction of the climate signal from glacier fluctuations. Therefore, glacier response time, size distribution, geographical and climatological location are to be considered when extrapolating fluctuation observations from a few glaciers or from an individual region to the entire Alpine glacierisation.

Annual front variations are relatively easy to measure in the field (as compared to mass balance measurements) and can be checked against decadal length change measurements from satellite images (cf. Kappeler 2006). In addition, this series have an important didactic value for the wider public, especially in places where information on glacier change is well documented and presented in the glacier forefield (e.g., Figure 5.5; Maisch et al. 1999). However, parallel with the dramatic mass loss of Alpine glaciers in the past two decades, geometries of many glaciers have become decoupled from current climate, and hence, glacier length change has definitely become a climate indicator with non-linear behaviour.

Figure 5.5: Glacier trail in the forefield of Morteratsch Glacier, Switzerland. Signs along the path towards the glacier terminus provide information about the glacier retreat and refer to related topics, described in a guide published by Maisch et al. (1999). The photo was taken in May 2005.

Mass balance measurements, providing a direct and undelayed glacier signal of annual atmospheric conditions, become even more important in times of fast-changing climate. In this context, profiles showing net balance vs altitude, as well as net balance vs (calculated) ELA and net balance vs (calculated) AAR are valuable (cf. IUGG(CCS)/UNEP/UNESCO/WMO 2005) and thus should be calculated (and published) for each glacier in the (direct glaciological) mass balance observation network. Winter and summer balances provide intra-annual climate information and should therefore be surveyed on all glaciers with mass balance observations (see also Dyurgerov & Meier 1999, Vincent 2002). In view of the large contribution of glaciers smaller than 1 km^2 to glacier shrinkage in the past and the predictions of ongoing global warming (e.g., IPCC 2001, Schär et al. 2004, Beniston 2005a, b), future work should include studies on the influence of atmospheric warming on small glaciers and on current downwasting processes (see also Paul et al. 2004). However, climatic sensitivity of glaciers not only depends on glacier size, but also on their sensitivity to regional climate variability versus local topographic effects, which potentially complicates the extraction of a regional or global climate signal from glacier fluctuations (Kuhn et al. 1985, Vincent et al. 2004). Mass balance and ice flow models calibrated with available fluctuation data are needed to quantify these effects (e.g., Oerlemans 1998, Oerlemans 2001, Paul et al. in press).

Reconstructed glacier states are needed in order to put the measured glacier fluctuations of the last 150 years into context with glacier variability of the Holocene (cf. *Chapter 3.4*). There exist reconstructed front variation series for several Alpine glaciers, spanning time periods from centuries to millennia (e.g., Nicolussi 1994, Pelfini 1999, Nicolussi and Patzelt 2000, Holzhauser & Zumbühl 2003, Holzhauser et al. 2005). These works, however, have not been prepared within an international framework and, hence, most of the data are not available in digital form to the public and the scientific community.

6 Conclusions

The overall aim of the present study, carried out within the framework of the EU-funded ALP-IMP project and the WGMS, was the investigation of glacier fluctuations in the entire European Alps between 1850 and the end of the 21st century. The study demonstrates the potential of new technologies (satellite imagery, DEM, geo-informatics) for using information from glacier monitoring (inventory and fluctuation data) in combination with numerical modelling for the quantitative assessment of past, present and future glacierisations of an entire mountain range. The main conclusions can be summarised as follows:

This study compiled an unprecedented data set containing inventory information on the about 5,150 Alpine glaciers, as well as front variation and mass balance information from more than 670 and 25 of these glaciers, respectively. Continuous front variation surveys in the Alps started around 1880, while direct mass balance measurements were initiated in 1948.

The spatio-temporal analysis of this data set showed an overall glacier area loss since 1850 of about 35% until the 1970s (with a total ice cover of 2,909 km^2) and almost 50% until 2000. Rapidly shrinking glacier areas, spectacular tongue retreats and increasing mass losses are clear signs of the atmospheric warming observed in the Alps in the last 150 years and its acceleration over the past two decades. Mass balance data from nine Alpine glaciers with continuous measurements since 1967 show a mean specific (annual) net balance of -0.07 m w.e. in the 1970s, -0.44 m w.e. in the 1980s and -0.77 m w.e. in the 1990s. The ice loss between 1967 (1981) and 2004 cumulated to almost 18 m w.e. with a maximum annual ice loss of 2.5 m w.e. and a corresponding estimated total glacier volume loss of 5–10% in the extraordinary year of 2003.

In addition, a numerical model was developed, based on an empirical relation between precipitation and temperature at the glacier ELA_0 derived from direct glaciological mass balance measurements. It enables the computation of the glacierisation (i.e., the $rcELA_0$ and the corresponding cAA) over the entire European Alps from gridded climatologies and a DEM. The model experiment shows that a change in 6-month summer temperature by ±1 °C would be compensated by a precipitation increase/decrease of 25%. A rise in 6-month summer temperature of 3 °C accompanied by an increase in annual precipitation of 10% would lead to a mean rise of the $rcELA_0$ of about 340 m and would reduce the cAA of 1971–1990 by about 75%. In the event of a 5 °C summer temperature increase, the Alps would become almost completely ice-free.

At the end of the LIA (around 1850) most Alpine glaciers reached their Holocene maximum extent. During the following general retreat, intermittent periods of glacier re-advances in the 1890s, 1920s and 1970–1980s can still be seen. The mass loss after 1980 continued at high, even accelerated rates, culminating in the extraordinary year 2003. Glacier shrinkage of the past two decades indicates transition towards rapid downwasting rather than dynamic glacier response to changing climate. Present glacierisation in the European Alps is about to pass the Holocene minimum extent. According to the presented impact study, based on the IPCC (2001) climate scenarios for the 21st century, the possibility of Alpine glaciers disappearing within decades is far from negligible.

7 Outlook

The present study demonstrates the potential of an integrative monitoring strategy, combining glacier inventories, in-situ measurements and numerical modelling for analysing past, present and future glacierisation of an entire mountain range. Future needs, recommendations and open questions arising from this work are briefly summarised in this chapter and cited according to their importance to (a) future research on Alpine glaciers, (b) international glacier monitoring, and (c) the wider public, decision-makers and politicians:

Future research on Alpine glaciers should focus on:

- the acquisition of new insights into the causes of regional glacier fluctuations by combining the new, extensive glacier data set with the recently available gridded climatologies and DEM of high spatial resolution (about 100 m);

- the identification and quantification of the contribution of individual climate parameters to the Alpine glacier retreat since 1850;

- the spatio-temporal analysis of winter, summer and annual net balance in comparison with the corresponding prevailing synoptic patterns;

- the application of the presented methods to, and comparison with, other mountain regions;

- the solution to the question of whether there is a plausible AAR_0 for Alpine glaciers, and if so, whether it remains constant over time;

- the quantification of the current decoupling of glacier geometries from steady-state conditions due to the fast warming since the mid-1980s;

- the investigation of the representativity of the Alpine mass balance measurements for the volume loss of the entire Alpine glacierisation; and

- the influence of projected climate change on small glaciers and on downwasting processes (in view of the large contribution of glaciers smaller than 1 km^2 to glacier shrinkage in the past).

- Concerning the modelling of distributed $rcELA_0$ and cAA, future work should continue by:
    - strengthening the relationship between precipitation and temperature at the ELA_0 by using a larger data sample from direct glaciological mass balance measurements, and by extending the data range towards more maritime and more continental glaciers;
    - using the model in the inverse sense for the reconstruction of one climate variable (precipitation or temperature) from reconstructed glacier ELA_0 and the other climate variable;
    - comparing the modelled $rcELA_0$ and corresponding cAA with results from distributed mass balance models, and the quantification of the influence of neglected feedback mechanisms and topographic effects; and
    - extending the model to consider other relevant parameters that can be derived from a DEM (e.g., potential solar radiation).

International glacier monitoring should:

- overcome national boundaries (successful glacier monitoring is organised on national levels, coordinated within the mountain regions and following international standards and strategies, as developed for instance by the GLIMS project and the WGMS);

- compile Alpine-wide glacier inventories for 1850 from available data and new reconstructions, and for 2000 (or 2003, due to clear sky conditions) from satellite images;

- link the available and new inventories on a glacier-by-glacier level (to ensure unity of samples when calculating glacier changes between the different inventories);

- increase the repetition rate of inventory activities from a few decades (as proposed earlier in international monitoring programmes) to about one decade, due to the current high rates of ice loss;

- maintain and revive long-term fluctuation series. In this way, decadal glacier length changes as derived from satellite images, and volume/thickness changes from repeated surveys using traditional (e.g., triangulation, photogrammetry) or new (e.g., laser scanning, interferometric synthetic aperture radar) technologies, could help to detect errors and systematic biases in front variation and mass balance series;

- integrate glacier reconstruction activities into the global framework of WGMS, to extend existing fluctuations series back in time; and

- better integrate modelling studies into the international monitoring strategies, as only with the help of an ensemble of modelling studies on different scales of complexity is it possible to simulate glacier changes of entire mountain ranges and at the same time to understand and quantify the physical processes governing glacier energy/mass balance and flow.

Wider public, decision-makers and politicians should be aware of the fact that:

- increasing mass loss, rapidly shrinking glacier areas, disintegrating and spectacular tongue retreats are clear signs of the atmospheric warming observed in the Alps during the last 150 years and the acceleration observed over the past two decades;

- however, in the short term or at a regional scale, glaciers show a highly individual variability (glacier behaviour depends not only on regional climate but also on local topographic effects which complicate the extraction of the climate signal from glacier fluctuations);

- hence, front variation from a few glaciers or fluctuations of a small regional glacier sample cannot per se be considered to be representative of the entire Alpine glacierisation;

- due to the strong mass loss during the past two decades, glacier geometries are in disequilibrium with the current climate and thus will continue to retreat during the next years or decades (depending on a glacier specific delay), even if mean air temperatures were halted at today's level;

- from a model experiment, the investigation of the impact of climate scenarios on glaciers leads to the conclusion that the possibility of the majority of Alpine glaciers disappearing within decades is far from slight and under the assumption of such scenarios, only the largest and highest-reaching Alpine glaciers could survive into the 22nd century; and

- especially in densely populated high mountain areas such as the European Alps, one should start immediately to consider the consequences (and implement possible counter-measures) of such extreme glacier wasting on the hydrological cycles, freshwater resources for agriculture and industry, hydro-power production, tourism, and natural hazards.

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Acknowledgements

"[His] study was a total mess, like the results of an explosion in a public library."

Douglas Adams (1979), fortunately referring to Slartibartfast's study.

The way to this PhD was for me like an expedition through an Alpine mountain range with fascinating, glaciated landscape, exhausting ascents, sometimes rewarded with great views or insights, sometimes followed by awkward descents. Fortunately, I met a lot of great people along the way who helped me to arrive, finally, at my destination. Though it is not possible to name all of them personally here, I will name the most outstanding, and their institutions.

My sincere thanks go, therefore, to:

- Wilfried Haeberli, who inspired me with his fascination for glaciers and fine-tuned my knowledge of glaciology. He initiated this study, indicated the direction to take and trusted me to discover my own route.

- Martin Hoelzle, for helping to prepare the expedition load and for accompanying me along the first stretch. With his abundant experience he was always a good discussion partner when I had to decide which valley to take.

- the Glaciology and Geomorphodynamics Group (3G) at the Department of Geography, University of Zurich, which is the best bunch of people you could ever want for fruitful scientific discussions, for support during harsh field work conditions, but also for a spirited social life.

- the organisers and lecturers at the Karthaus Summer School on *Ice Sheets and Glaciers in the Climate System* in 2003, for opening my mind to the glaciological undertakings beyond those accessible at my own institute.

- the organisers and lecturers at the NCCR Summer School in Grindelwald in 2005, with the topic *From the Holocene to the Anthropocene: Climate of the Last 1000 Years*. They helped me to bring my findings in glacier research into context with the state-of-the-art of the current scientific climate change discussion.

- the partners of the ALP-IMP project for the friendly integration into their community and the great collaboration, especially with the Central Institute for Meteorology and Geodynamics in Vienna, Austria.

- all national correspondents and principal investigators of WGMS as well as to all other researchers and institutes that have been providing data to WGMS for many years.

- the Academia Engiadina, as sub-contractor of ALP-IMP, and the numerous student trainees of WGMS for their contribution to the control, correction and formatting of the enormous data load.

- my colleagues from the WGMS for their daily efforts and the exemplary teamwork.

- Kaspar Kundert and colleagues from ESRI Geoinformatik AG in Zurich. Working for this company allowed me to keep my GIS knowledge at a top level and to actively profit from their experience in time and project management.

- Susan Braun-Clarke for carefully editing the English, and Regula Frauenfelder for providing me with her thesis (Frauenfelder 2005) as the model for structure and layout of the present study.

- my parents, my sis, and *the blokes* for helping me not to forget the non-frozen things in life.

- Gabriela, for having joined me on a large part of this way.

The present study was funded by the Swiss Federal Office of Education and Science (BBW-contract: 901.0498-2) within the EU programme ALP-IMP (contract EVK2-CT-2002-00148).

Part B: Papers

Paper I

Orlove, B., Wiegandt, E. and B. Luckman (eds.):

The darkening peaks: Glacial retreat in scientific and social context

Glacier fluctuations in the European Alps 1850–2000: an overview and spatio-temporal analysis of available data

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University of Zurich, Switzerland

Abstract: Within the framework of the EU-funded project ALP-IMP, which deals with climate change within the greater Alpine[1] region, the World Glacier Monitoring Service (WGMS) compiled an unprecedented Alpine glacier dataset containing fluctuation information dating back to 1850. Here we give an overview of the available data, results of spatio-temporal analyses and discuss the representativity of front variation and mass balance series for the entire Alpine glacierisation. In the 1970's there were approximately 5,150 Alpine glaciers covering an area of 2,909 km^2. Overall area loss since 1850 is estimated to be about 35% until the 1970's and almost 50% until 2000. Rapidly shrinking glacier areas, spectacular tongue retreats and increasing mass losses are clear signs of the atmospheric warming observed in the Alps in the last 150 years and its acceleration over the past two decades. However, in the short-term or at a regional scale, glaciers show a highly individual variability. Glacier behaviour depends not only on regional climate but also on local topographic effects which complicate the extraction of the climate signal from glacier fluctuations. The latter are essential for the verification of mass balance and ice flow models, which are needed to quantify these topographic effects. It is of major importance to continue with long-term fluctuation measurements and to extend the existing fluctuation series back in time using reconstructions of former glacier geometries. Additionally, it is necessary to integrate glacier monitoring as well as reconstruction activities into the global framework of the Glacier Land Ice Measurement from Space (GLIMS) project and the WGMS.

1 Introduction

Fluctuations of mountain glaciers are among the best natural indicators of climate change (IPCC 2001). Changes in precipitation and wind lead to variations in accumulation, while changes in temperature, radiation fluxes and wind, among other factors, affect the surface energy balance and thus ablation. Disturbances in glacier mass balance, in turn, alter the flow regime and, consequently, after a glacier specific delay, result in a glacier advance or retreat such that the glacier geometry and altitude range change until accumulation equals ablation (Kuhn et al. 1985). Hence, mass balance is the direct and undelayed signal of annual atmospheric conditions, whereas changes in length are an indirect, delayed, filtered but also enhanced signal (Haeberli 1998).

The modern concept of worldwide glacier observation is an integrated and multi-level one, it aims to combine in-situ observations with remotely sensed data, process understanding with global coverage and traditional measurements with new technologies. Such concept of global glacier observation uses detailed mass and energy balance studies from just a few glaciers together with length-change observations from many sites and inventories covering entire mountain chains. Numerical models link all three components over time and space (Haeberli 2004). The EU-funded project **ALP-IMP** is an integrated research that focuses on multi-centennial climate variability in the **ALP**s based on **I**nstrumental data, **M**odel simulations and **P**roxy data. It represents a unique opportunity to apply this

[1] In this article 'Alps' or 'Alpine' refer explicitly to the European Alps, the terms 'alps' or 'alpine' are purely generic.

glacier monitoring concept to the European Alps where, by far, the most concentrated amount of information about glacier fluctuations over the last century is available. The World Glacier Monitoring Service (WGMS) has compiled, within the framework of the ALP-IMP project, an unprecedented dataset containing inventory data (i.e., area, length and altitude range) from approximately 5,150 Alpine glaciers and fluctuation series from more than 670 of them (i.e., more than 25,350 annual front variation and 575 annual mass balance observations) dating back to 1850.

In this study we give an overview of available glacier datasets from the European Alps and analyse glacier fluctuations between 1850 and 2000. To achieve this, glacier-size characteristics from the 1970’s are analysed, this is the only time period for which a complete Alpine inventory is available. From this inventory, Alpine glacierisation in 1850 and in 2000 are extrapolated using size-dependent area changes from Switzerland. Furthermore, mass balance and front variation series are investigated to provide a better insight into glacier fluctuations, corresponding acceleration trends and regional distribution patterns at an annual resolution. Finally, the representativity of these recorded fluctuation series for the entire Alpine glacierisation is discussed and conclusions for glacier monitoring are drawn.

This article is structured as follows: A brief summary of the history of international coordinated glacier observations and the most relevant references are given in the section ‘Background’. The third section ‘Data’ describes the different types and spatio-temporal distribution of available glacier information from the European Alps. The section ‘Analysis and results’ presents the size characteristics of Alpine glacierisation in the 1970’s, extrapolates the total glacier area in 1850 and 2000 and analyses front variation and mass balance observations. The fifth section discusses spatial data coverage and glacier shrinkage over the past 150 years as well as the representativity of the Swiss glacier inventory and fluctuation series (mass balance and front variation) for the entire Alpine glacierisation. Section six sums up and concludes the main results of this study.

2 Background

A worldwide collection of information about ongoing glacier changes was initiated in 1894 with the foundation of the International Glacier Commission at the 6th International Geological Congress in Zurich, Switzerland. At that time, the Swiss limnologist F.A. Forel started publishing the periodical 'Rapports sur les variations périodiques des glaciers' on behalf of the then established 'Commission Internationale des Glaciers' (Forel 1895). Up until 1961, the data compilations constituting the main source of length-change data worldwide were published in French, Italian, German and English. Since 1967, the publications have all been written in English. The first reports contain mainly qualitative observations with the exception of the glaciers in the European Alps and Scandinavia, many of which have had extensive documentation and quantitative measurements recorded from the very beginning. After the First World War, P.L. Mercanton edited the publications which began to appear less than annually. Since 1933 they were published on behalf of the International Commission on Snow and Ice (ICSI), a part of the International Association of Hydrological Sciences (IAHS). From 1967 the data were published in five-year intervals under the title 'Fluctuations of Glaciers', at first by the Permanent Service on the Fluctuations of Glaciers (PSFG, Kasser 1970) then – after the merger of PSFG with the Temporary Technical Secretariat for the World Glacier Inventory (TTS/WGI) in 1986 – by the World Glacier Monitoring Service (WGMS). An extensive overview of the corresponding literature is given in Hoelzle et al. (2003). The need for a worldwide inventory of perennial snow and ice masses was first considered during the International Hydrological Decade operated by the United Nations Educational, Scientific and Cultural Organisation (UNESCO) from 1965 until 1974 (UNEP/GEMS 1992). Preliminary results and a thorough discussion of the techniques and standards applied for glacier inventorying was given in IAHS (1980). The status report and the corresponding national literature of all national glacier inventories compiled at that time was published in WGI (1989). More detailed results of glacier area changes for specific regions or countries can be found in the literature, often with special emphasis on developments since 1850. Some examples for the European Alps are CGI–CNR (1962) for Italy, Gross (1988) for Austria and Maisch et al. (2000) for Switzerland as well as Vivian (1975) for the Western Alps, Maisch (1992) for the Grisons (CH), Böhm (1993) for the Goldberg region (Hohe Tauern, Austria) and Damm (1998) for the Rieserferner group (Tirol, AT).

3 Data

In the following chapter the three types of available Alpine glacier information are described. The geographical distribution of the different datasets is shown in Figure 1.

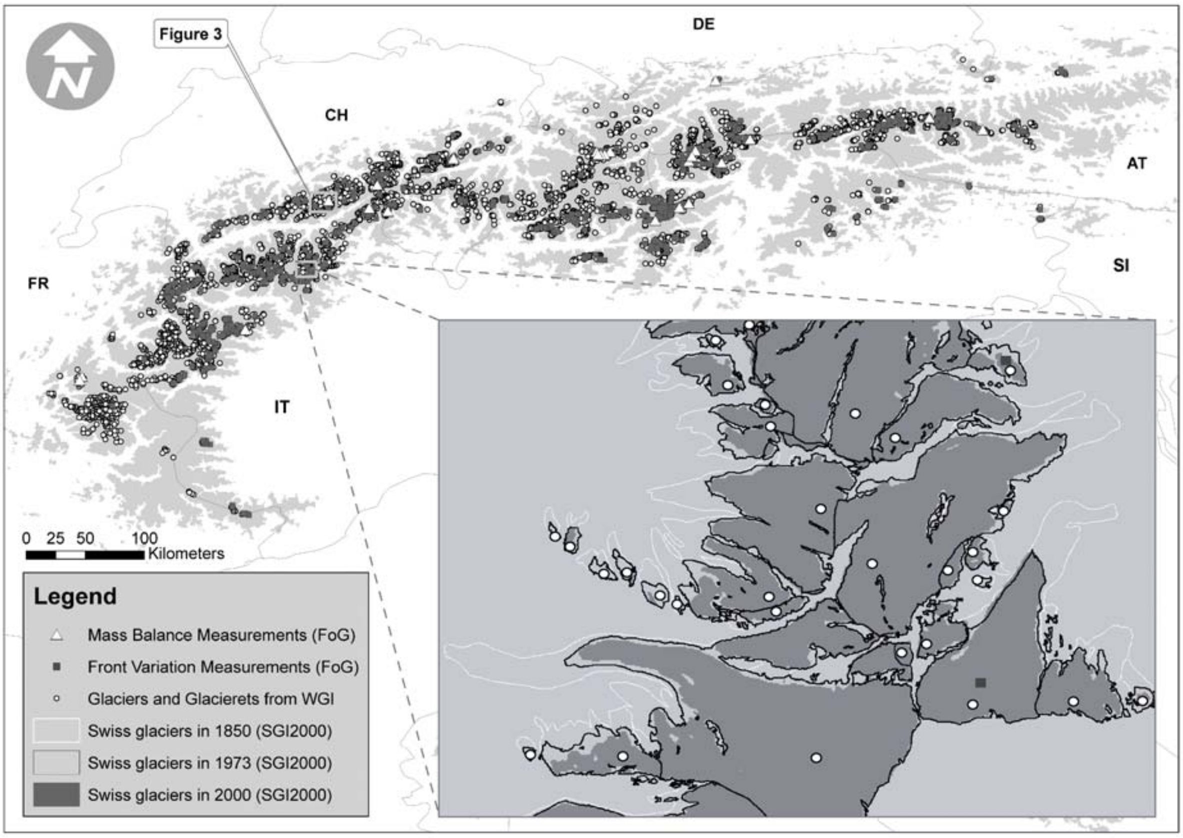

Figure 1: Geographical distribution of available glacier information in the Alps: WGI data (white circles) as well as mass balance data (white triangles) and front variation data (dark grey squares) from the FoG database. Elevations above 1,500 m a.s.l. are in light grey. Country names are abbreviated as follows: Austria (AT), France (FR), Germany (DE), Italy (IT), Slovenia (SI) and Switzerland (CH). The inset shows glacier polygons of 1850, 1973 and 2000 from the SGI2000.

World Glacier Inventory (WGI)

The WGI contains attribute data on glacier area, length, orientation and elevation as well as a classification of morphological types and moraines, linked to the glacier coordinates (Fig. 1, WGI). The inventory entries are based upon a specific observation time and can be viewed as a snapshot of the spatial glacier distribution. The data are stored in the WGI database (part of WGMS database) and are published in the WGI (1989). The latter summarises national inventories for the entire Alps.

Complete national inventories for the European Alps are available for Austria (1969), France (1967–71), Switzerland (1973), Germany (1979) and Italy (1975–84). The inventories of Austria, Switzerland and Germany refer to one single reference year while the records of France and Italy are compiled over a longer period of time to achieve total coverage (Figure 2). However, in all inventories there is a certain percentage of glaciers for which no data from the corresponding reference period/year could be obtained so information from earlier years has been substituted. For example, in the Swiss inventory only data from 1,550 glaciers date from 1973 while the information from the remaining 274 glaciers refers to earlier years (Figure 2, CH (WGI)). Glacier identification, assignment and partitioning (due to glacier shrinkage) are the main challenges for comparisons of inventories overlapping in space or time. Therefore, the total number and areas of glaciers may vary in different studies. The WGI (1989) sums the total area of the 5,154 Alpine glaciers from Austria (542 km^2), France (417 km^2), Switzerland (1342 km^2), Germany (1 km^2) and Italy (607 km^2) to be 2,909 km^2. Due to previously mentioned inconsistencies, the dataset used in this study differs slightly from these numbers. The Italian Inventory (1975–84) sums up to only 602 km^2 and the total number of Alpine glaciers to 5,167, i.e. 13

glaciers more than in the original status report. These differences, however, are smaller than 0.3% and therefore negligible.

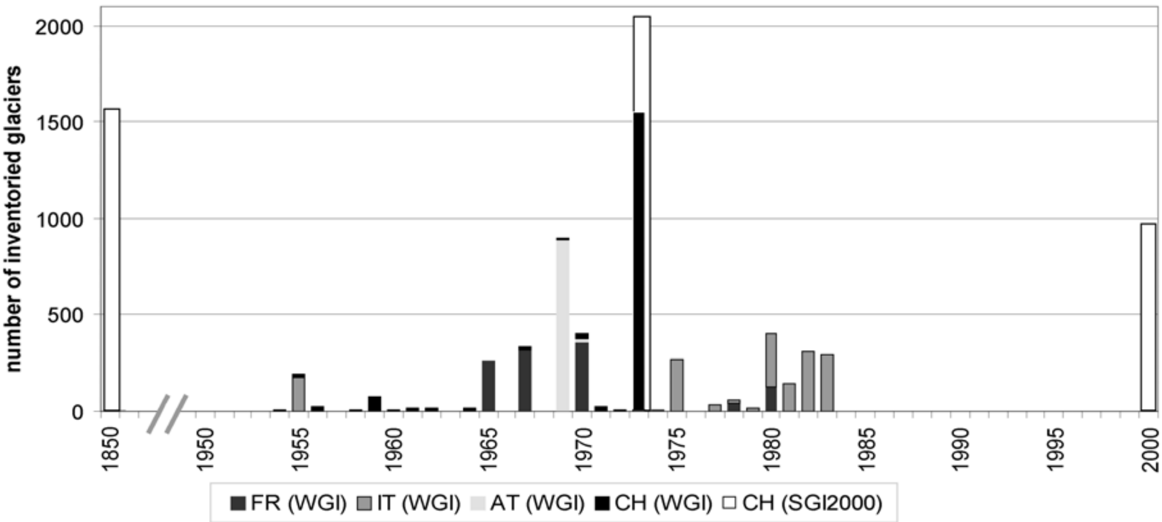

Figure 2: Frequency of glacier inventories in the Alps. Number of inventoried glaciers per year are marked according to country and data source (WGI and SGI2000). For example, in 1973 there are inventory data in the WGI from six Italian, two Austrian and 1550 Swiss glaciers, while in the SGI2000 there is information on 2,057 Swiss glaciers.

Swiss Glacier Inventory 2000 (SGI2000)

The new SGI2000 has been compiled from multispectral Landsat Thematic Mapper (TM) data acquired in 1998–1999 (path-row 194/5-27/8). Glacier information (e.g. area, slope, aspect, equilibrium line altitude) was obtained from a combination of glacier outlines with a digital elevation model and the related statistical analysis by a Geographic Information System (Kääb et al. 2002, Paul et al. 2002, Paul 2004). Several glaciers were not properly identified due to cast shadow, snow cover and debris and were excluded from the statistical analysis. Altogether, new areas for 938 glaciers were obtained for 2000 (Figure 2) and the related topographical information extracted. The former glacier inventories from 1850 and 1973 were digitised from the original topographic maps and are now a major part of the SGI2000 (Figure 3). The 1973 outlines are also used to define the hydrological basins of individual glaciers in the satellite derived inventory, in particular the ice-ice divides. However, due to different identification codes used in the inventories of Müller et al. (1976), Maisch et al. (2000) and the SGI2000, a direct comparison of glacier areas is not yet possible. Moreover, glacier retreat has caused severe changes in glacier geometry (tongue separation, disintegration, etc.), which prevents a direct comparison. For this reason our analysis of glacier changes was based on different samples (Table 2). Major results of this study were summarised in Paul et al. (2004) and in Table 2.

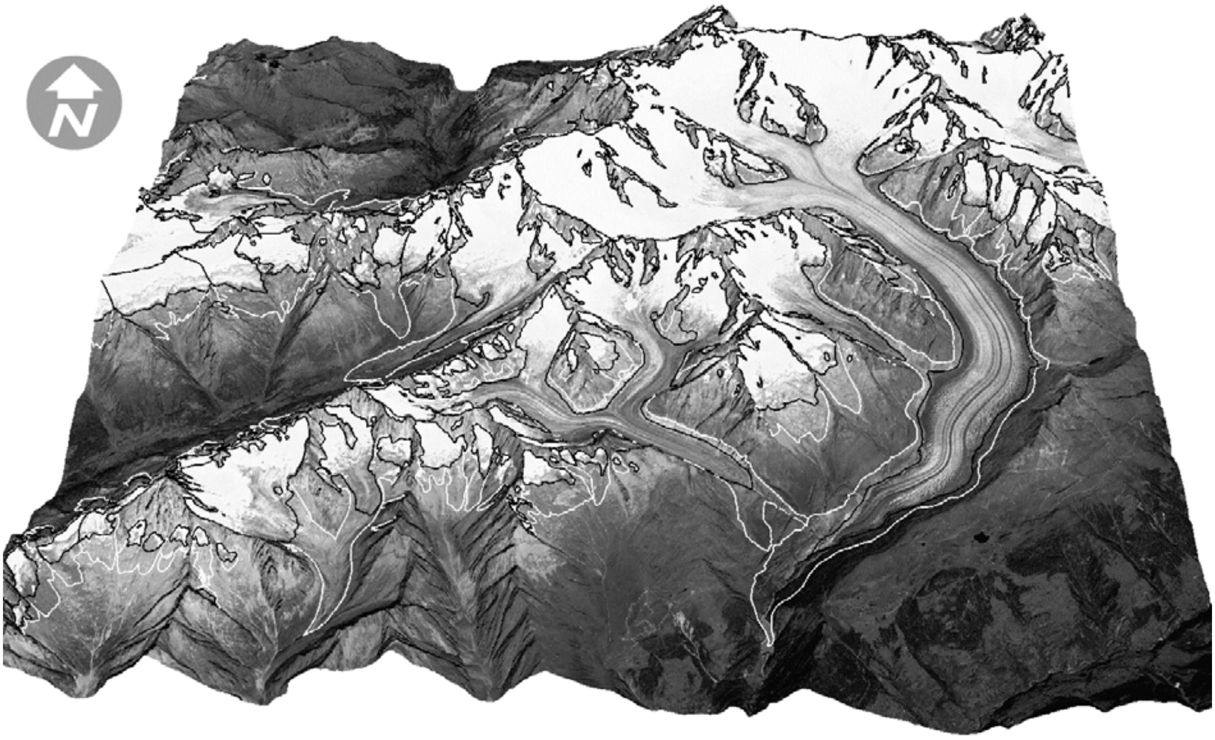

Figure 3: Synthetic oblique-perspective of the Aletsch glacier region, Switzerland, generated from a digital elevation model (DEM25; reproduced by permission of swisstopo, BA057338) overlaid with a fusion of satellite images from Landsat TM (1999) and IRS-1C (1997) in a greyscale rendition. The Grosser Aletsch glacier retreated about 2,550 m from 1850 (white lines) to 1973 (black lines) and another 680 m by 2000.

Fluctuations of Glaciers (FoG)

The FoG database contains attribute data on glacier changes over time, i.e. front variations, mass balance, changes in area, thickness and volume, linked to glacier coordinates (Figure 1, FoG). The data are stored in the FoG database (part of WGMS database) and published in the Fluctuations of Glaciers series on 5-year intervals (latest edition: FoG 2005), and biannually in the Glacier Mass Balance Bulletin (latest edition: GMBB 2005).

Regular glacier front variation surveys in the Alps started around 1880. The number of glaciers surveyed and the continuity of series changed over time due to world history and the perceptions of the glaciological community (Haeberli and Zumbühl 2003, Haeberli this issue). Direct measurements of glacier mass balance in the Alps started at Limmern (CH) and Plattalva (CH) in 1948, followed by Sarennes (FR) in 1949, Hintereis (AT) and Kesselwand (AT) in 1953 with others following in successive years. In the last reporting period (1995–2000) 297 glacier front measurements were made along with the mass balance of 18 Alpine glaciers (FoG 2005). For the analysis in chapter 3 only front variation series with more than nine survey years and mass balance series longer than three years have been considered.

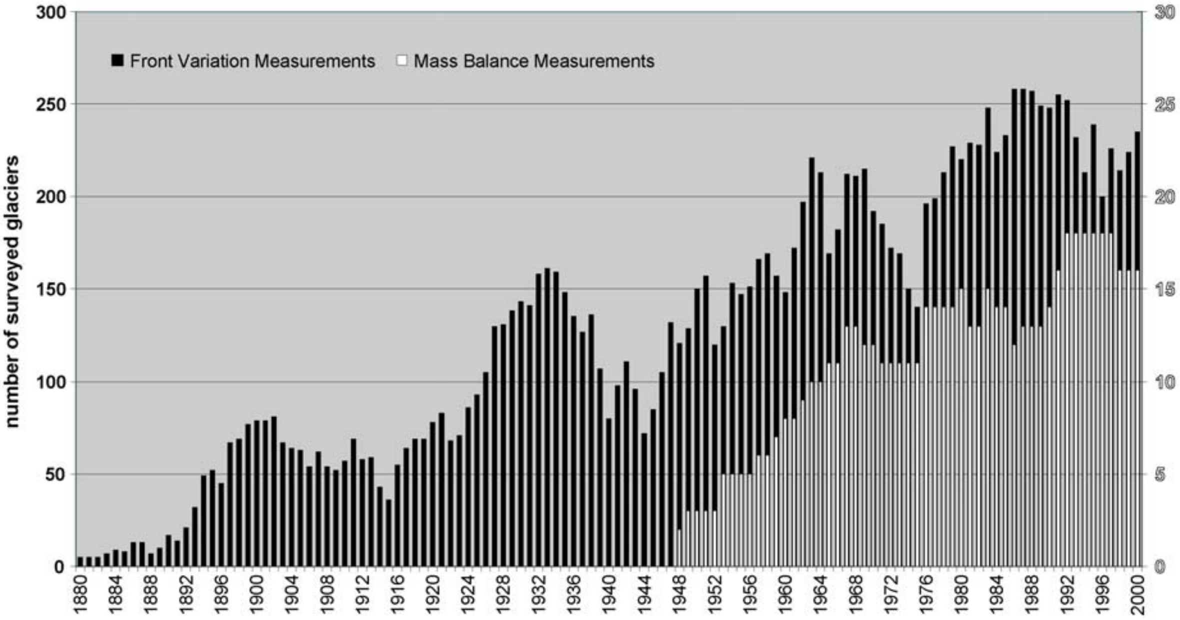

Figure 4: Frequency of front variation and mass balance measurements in the Alps. The number of glaciers with front variation (black bars, left axis) and mass balance (white bars, right axis) observations are shown from 1880–2000. Only glaciers with more than 18 front variations or three mass balance surveys are considered.

There are some reconstructed front variation series for several Alpine glaciers, spanning time periods from centuries to millennia (e.g. Holzhauser and Zumbühl 1996, Holzhauser 1997, Nicolussi and Patzelt 2000, Holzhauser et al. 2005). In addition, there are some studies that estimate secular mass balance trends from cumulative glacier length changes (e.g. Haeberli and Holzhauser 2003, Hoelzle et al. 2003) or from glacier surfaces reconstructed from historical maps (cf. Haeberli 1998, Steiner et al. this issue). These works, however, have not been prepared within an international framework and most of the data are not publicly available, so we have not considered them here.

4 Analysis and results

Alpine glacierisation in the 1970's

The only complete Alpine inventory available is from the 1970's, with 5,154 glaciers and an area of 2,909 km^2 (WGI 1989). Paul et al. (2004) estimated the total ice volume to be about 100 km^3, much lower than the 130 km^3 suggested earlier by Haeberli and Hoelzle (1995). The latter estimated the total ice volume from the total Alpine glacier area and an averaged thickness from all the glaciers (in accordance with semi-elliptical cross-sectional glacier geometry). Paul et al. (2004) calculated the total volume loss (-25 km^3) for the period 1973–1998/99 from the mean Alpine glacier area (2,753 km^2) and the average cumulative mass balance for eight Alpine glaciers (-9 m w.e.). Assuming that the relative change in volume is likely to have been larger than the corresponding relative change in area (for geometrical reasons), then the estimated relative volume loss is roughly -25% and, therefore, the total Alpine ice volume in the 1970's was about 100 km^3.

Figure 5 shows the number and area of Alpine glaciers (AT, CH, FR and IT) according to glacier size classes. 82% of the glaciers are smaller than 0.5 km^2 and cover 21% of the total glaciated area. Glacierets and névés (perennial snow banks) don't normally show dynamic reactions and are, therefore, usually excluded from glacier studies. However, neglecting these small glaciers in inventories could introduce significant errors in the assessment of regional glacier change. Only seven glaciers (Grosser Aletsch (CH), Gorner (CH), Fiescher (CH), Mer de Glace (FR), Unteraar (CH), Unterer Grindelwald (CH) and Oberaletsch (CH)) are larger than 20 km^2, but represent 10% of the total area. Glaciers between 1 and 10 km^2 account for 46% of the Alpine glacier area.

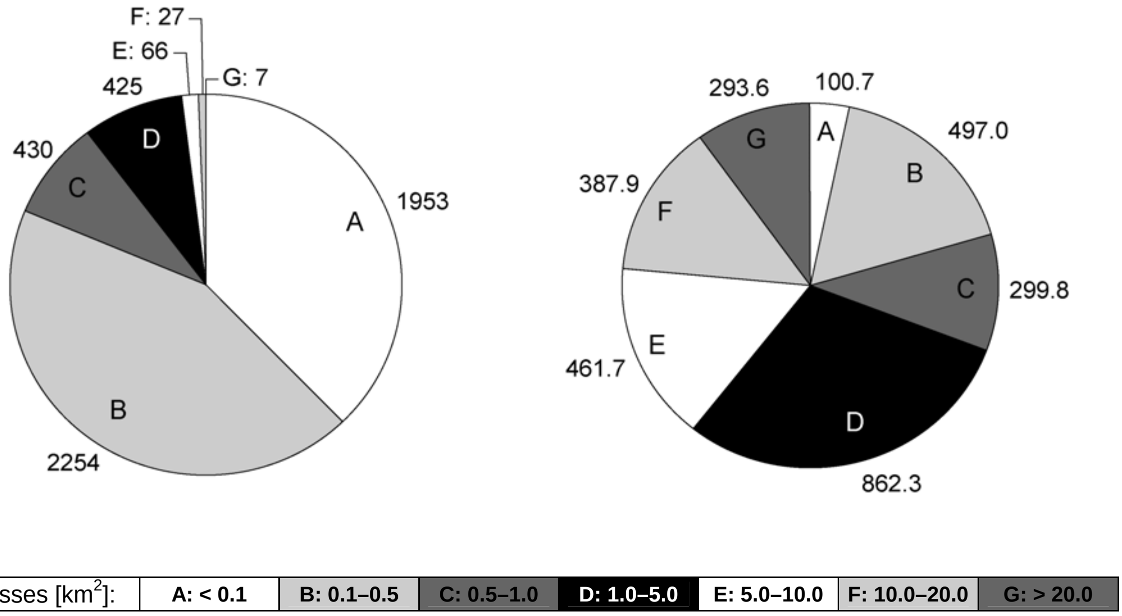

Figure 5: Distribution of glaciers by number and size in the Alps for the 1970's. Pie charts give percentages by number (left) and area (km^2, right) with absolute values indicated. In total the 5162 glaciers cover an area of 2902.9 km^2 (Table 1a, Alps WGI). The five German and two Slovenian glaciers are not considered in this figure.

From Table 1a the regional distribution of number and area of Alpine glaciers can be calculated for each Alpine country. Most of the glaciers are located in Switzerland (35%), followed by Italy (27%), France (20%) and Austria (18%). Regarding total glacier area, the majority of European ice is located in Switzerland (46%) and Italy (21%). Due to several large glaciers, Austria ranks third with 19% of the Alpine glacier area, followed by France with 14%. The five German glaciers with a total area of 1 km^2 and the two small Slovenian glaciers are not considered in Tables 1 and 2.

Tables 1a and b show the glacier size characteristics of the Austrian, French, Italian and Swiss glacierisations in the 1970's. The number of glaciers in each area class is very similar in all countries except for France where 50% of the glaciers are smaller than 0.1 km^2. The area distribution in Austria and Italy is dominated equally by small and mid-size glaciers. Mer de Glace (FR) with an area of 33 km^2 corresponds to almost 8% of the French glacierisation. In Switzerland the 22 largest glaciers (> 10 km^2) account for 37% of the total glacier area.

Table 1: Distribution of glaciers by number and size in the Alps in the 1970's. Absolute values (a) and percentage (b) are given in seven size classes for glaciers in the entire Alps, Austria (AT), Switzerland (CH), France (FR) and Italy (IT), as well as for surveys of Alpine glaciers with more than nine measurements of front variation (FV), and for surveys with more than three mass balance measurements (MB). Number and area are shown in separate columns to facilitate comparison between the different data samples and size classes.

a) number and area in absolute values

classes [km^2]		**Alps** WGI	**AT** WGI	**CH** WGI	**FR** WGI	**IT** WGI	**Alps** FoG, FV	**Alps** FoG, MB
0.0–0.1	number	1953	287	636	522	508	16	0
	area [km^2]	100.7	16.2	29.4	24.5	30.5	1.0	0.0
0.1–0.5	number	2254	416	826	361	651	130	3
	area [km^2]	497.0	92.3	185.5	77.0	142.2	36.2	0.9
0.5–1	number	430	112	156	73	89	92	4
	area [km^2]	299.8	77.6	108.4	51.4	62.5	64.0	2.8
1–5	number	425	95	152	79	99	198	13
	area [km^2]	862.3	213.2	294.8	153.6	200.7	446.6	35.2
5–10	number	66	10	32	7	17	56	3
	area [km^2]	461.7	71.8	223.0	51.0	115.9	392.1	24.6
10–20	number	27	5	16	2	4	27	2
	area [km^2]	387.9	71.2	240.1	26.1	50.5	396.0	33.0
>20	number	7	0	6	1	0	7	0
	area [km^2]	293.6	0.0	260.5	33.1	0.0	293.5	0.0
Total	number	5162	925	1824	1045	1368	526	25
	area [km^2]	2902.9	542.2	1341.7	416.6	602.4	1629.3	96.5

b) number and area in percent

classes [km^2]		**Alps** WGI	**AT** WGI	**CH** WGI	**FR** WGI	**IT** WGI	**Alps** FoG, FV	**Alps** FoG, MB
0.0–0.1	number [%]	37.8	31.0	34.9	50.0	37.1	3.0	0.0
	area [%]	3.5	3.0	2.2	5.9	5.1	0.1	0.0
0.1–0.5	number [%]	43.7	45.0	45.3	34.5	47.6	24.7	12.0
	area [%]	17.1	17.0	13.8	18.5	23.6	2.2	0.9
0.5–1	number [%]	8.3	12.1	8.6	7.0	6.5	17.5	16.0
	area [%]	10.3	14.3	8.1	12.3	10.4	3.9	2.9
1–5	number [%]	8.2	10.3	8.3	7.6	7.2	37.6	52.0
	area [%]	29.7	39.3	22.0	36.9	33.3	27.4	36.5
5–10	number [%]	1.3	1.1	1.8	0.7	1.2	10.6	12.0
	area [%]	15.9	13.2	16.6	12.2	19.2	24.1	25.5
10–20	number [%]	0.5	0.5	0.9	0.2	0.3	5.1	8.0
	area [%]	13.4	13.1	17.9	6.3	8.4	24.3	34.2
>20	number [%]	0.1	0.0	0.3	0.1	0.0	1.3	0.0
	area [%]	10.1	0.0	19.4	7.9	0.0	18.0	0.0
Total	number [%]	100	100	100	100	100	100	100
	area [%]	100	100	100	100	100	100	100

Alpine glacierisation in 1850 and 2000

Based on the Alpine inventory of the 1970's, the Alpine glacier area in 1850 and 2000 can be extrapolated by applying the relative area changes (1850–1973, 1973–2000) of the seven glacier-size classes from the SGI2000 to the corresponding Alpine glacier areas in the 1970's (Table 2). The estimated Alpine glacier areas amount to 4,474 km^2 in 1850, and to 2,272 km^2 in 2000. This corresponds to an overall glacier area loss from 1850 up until the 1970's of 35% and almost 50% by 2000 – or an area reduction of 22% between the 1970's and 2000. Dividing the total area loss by time provides estimates of area change per decade of 2.9% between 1850–1973 and of 8.2% between 1973–2000.

Table 2: Alpine glacierisation 1850, 1970's and 2000. [§]Alpine glacier area in 1850 and 2000 is extrapolated from the glacier area in the 1970's (WGI) and relative area changes of the seven glacier size classes in Switzerland (SGI2000). [*]The relative area changes in Switzerland are calculated from the comparable sub-samples, i.e. 1567 glaciers for 1850–1973 and 938 glaciers for 1973–2000, respectively.

	Switzerland (SGI2000)								Alps			
class	1850		1973		2000		1850–1973*	1973–2000*	1970's		1850§	2000§
[km^2]	num.	area [km^2]	num.	area [km^2]	num.	area [km^2]	area change [%]	area change [%]	num.	area [km^2]	area [km^2]	area [km^2]
< 0.1	297	17.3	1022	40.1	164	3.6	-55.4	-64.6	1953	100.7	225.5	35.6
0.1–.5	715	181.3	673	153.9	448	60.3	-52.9	-45.6	2254	497.0	1055.0	270.4
0.5–1	249	172.5	151	104.1	131	63.5	-44.3	-29.1	430	299.8	538.0	212.6
1–5	253	524.4	157	296.0	141	217.1	-33.2	-17.9	425	862.3	1291.1	707.9
5–10	26	195.5	35	249.4	36	232.6	-19.7	-10.8	66	461.7	574.8	412.1
10–20	18	259.9	14	216.3	13	192.8	-14.8	-8.2	27	387.9	455.1	356.1
> 20	9	270.5	5	225.9	5	213.0	-12.3	-5.7	7	293.6	334.8	276.9
Total	1567	1621.4	2057	1285.7	938	982.9	-27.1	-16.1	5162	2902.9	4474.3	2271.6

Several methods exist to calculate glacier volume from other variables, they are either based on statistical relationships (e.g. Müller et al. 1976), empirical studies (e.g. Maisch et al. 2000) or physical parameters (e.g. Haeberli and Hoelzle 1995). However, all of them apply glacier size as a scaling factor and the deviations between individual methods are large. As the individual glacier sizes for the year 2000 are not available yet for all glaciers, we have not attempted to present glacier volume evolution with time. However, a current estimate of Alpine glacier volume in 2000 indicates approximately 75 km^3 remains (Paul et al. 2004).

Alpine front variations

Large valley glaciers retreated continuously since the Little Ice Age maximum around 1850. Smaller mountain glaciers show marked periods of intermittent advances in the 1890's, 1920's and 1970–1980's. Front variations of smallest glaciers have a high annual variability. In Figure 6 front variation series with more than 18 measurement years are plotted and sorted according to glacier size. The advance periods of the 1920's and 1970–1980's as well as the retreat periods in-between and the one after 1990 show up very clearly. However, on the individual level the climate signal from variations in frontal position of glaciers is much more complex. This noise prevails even when the dataset is sorted according to response time (see Johannesson et al. 1989, and Haeberli and Hoelzle 1995) or analysed in geographical sub-samples. Figure 6 is dominated by smaller mountain glaciers (Table 1, Alps FoG, FV), therefore the signals of large valley glaciers and the smallest glaciers (including absolute retreat values) are more visible in the graphs of individual cumulative front variation (e.g. Haeberli et al. 1989, Hoelzle et al. 2003).

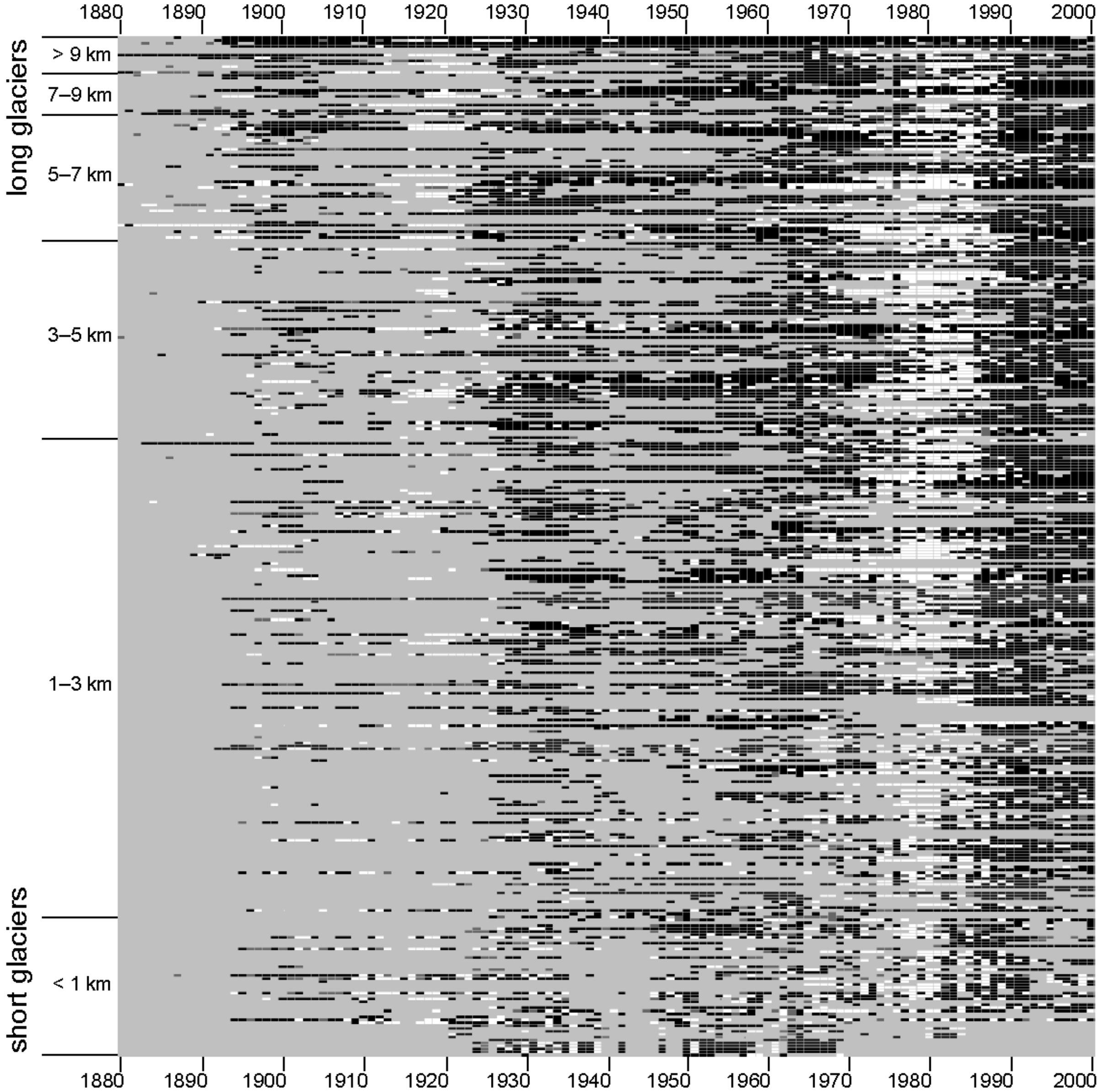

Figure 6: Alpine front variation series 1880–2000. Annual front variation values from glaciers with more than 18 measurements are coloured white after an advance, black after a retreat, dark grey - no apparent variation and light grey - no data. Each row represents one glacier. The glaciers are sorted after glacier length in the 1970's (y-axis).

Alpine mass balances

50 years of direct mass balance measurements show a clear trend of mass loss. Even though some of the measured glaciers gained mass during the 1960's to 1980's, ice loss has actually accelerated in the past two decades (Figures 7 and 8). With respect to the geographical distribution, years with a uniformly positive (e.g. 1965, 1977, 1978) or negative (e.g. 1964, 1973, 1983) Alpine mass balance signal, as well as years with a clear spatial gradient in net balance (e.g. 1963, 1976) or with heterogeneous signals can be mainly found before 1986. After 1981 uniformly negative mass balance years dominate.

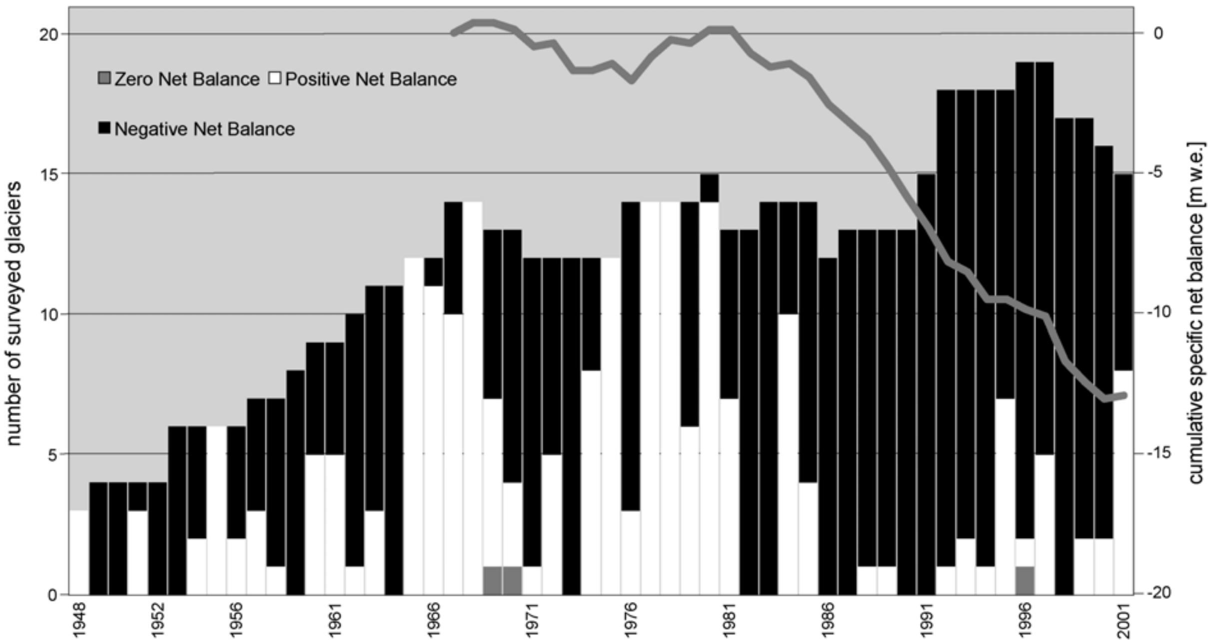

Figure 7: Alpine mass balance measurements from 1948–2001. Annual numbers of glaciers (left axis) with a zero net balance (dark grey), positive net balance (white), or negative net balance (black). In addition the graph of the mean cumulative specific net balance of the nine Alpine reference glaciers (Careser (IT), Gries (CH), Hintereis (AT), Kesselwand (AT), Saint Sorlin (F), Sarennes (F), Silvretta (CH), Sonnblick (AT) and Vernagt (AT)) is drawn from 1967–2001 (right axis).

Nine Alpine reference glaciers with continuous mass balance series over more than 30 years show a mean annual loss of ice thickness close to 37 cm w.e. per year, resulting in a total thickness reduction of about 13 m w.e. since 1967. The corresponding values for the period 1980–1999 are 60 cm w.e. per year and 12.3 m w.e., respectively (Table 3).

Table 3: Mean specific (annual) net balance of the Alpine reference glaciers: Careser (IT), Gries (CH), Hintereis (AT), Kesselwand (AT), Saint Sorlin (F), Sarennes (F), Silvretta (CH), Sonnblick (AT) and Vernagt (AT).

time period	number of reference glaciers	mean specific (annual) net balance [mm w.e.]
1950–1959	1–5	-536
1960–1969	6–9	-26
1970–1979	9	-69
1980–1989	9	-437
1990–1999	9	-767
1949–2001	1–9	-412
1967–2001	9	-369

5 Discussion

Data coverage

Glacier studies have a long tradition in the Alps that began with the establishment of systematic observation networks in the 1890’s (Haeberli this issue). In comparison to the rest of the world, the European Alps have the densest and most complete spatial glacier inventory through time (WGI 1989). Thus, the inventory data (WGI, SGI2000) contain information on spatial glacier distribution at certain times, whereas the fluctuations series (FoG) provide high resolution temporal information for specific locations. Interestingly, the 1970’s is the only period where an Alpine inventory with total spatial coverage can be compiled as most glaciers were relatively close to steady-state conditions (Figure 7, Patzelt 1985). The reconstructed glacier extents at the end of the Little Ice Age (around 1850) and the glacier outlines derived from multispectral satellite data around 2000 from the SGI2000 cover the major parts and the full range of area classes of Swiss glacierisation (Table 2). Thus, they can be used to extrapolate Alpine glacierisation in 1850 and 2000, based on the assumption that the relative losses of the different area classes in Switzerland are representative of other Alpine countries as well. This, of course, is not necessarily the case. The fluctuation series are numerous and well distributed over the Alps, with a minimum number of front variation series in the south-western part of the Alps (Figure 1, FoG). For the fluctuation series, length and completeness of the time series are most relevant.

Glacier shrinkage

The inventory of the 1970's and the extrapolated area estimates for 1850 and 2000 show the dramatic dimension of glacier shrinkage in the Alps. Despite the high degree of variability in individual glaciers, the European Alps have experienced a 50% decrease in ice coverage over the last 150 years. The area loss over each decade (in percent) between the 1970's and 2000 is almost three times faster than the loss of ice between 1850 and the 1970's. Variations in glacier frontal position provide a higher time resolution of the glacier retreat over the past 150 years. Though glaciers have generally been retreating since 1850, there have been several periods of documented readvances in the 1890's, 1920's and 1970–1980's (Patzelt 1985, Müller 1988, Pelfini and Smiraglia 1988, Reynaud 1988 and Haeberli et al. 1989). The area reduction after the 1970's mainly occurred after 1985 (see also, Paul et al. 2004). Thus, the acceleration of the glacier retreat in the last two decades was even more pronounced. Mass balance measurements are only available for the last five decades and confirm the general trend of glacier shrinkage. While some glaciers gained mass between 1960 and 1980, ice loss has accelerated over the last two decades. The mean specific (annual) net balance of the 1980's is 18% below the average of 1967–2001, and the value of the 1990's even doubles the average ice loss of 1967–2001. Most recent mass balance data show a continuation of the acceleration trend after 2000 with a peak in the extraordinary year of 2003 where the ice loss of the nine Alpine reference glaciers was about 2.5 m w.e. – exceeding the average of 1967–2000 by almost a factor of seven. Estimated total glacier-volume loss in the Alps in 2003 corresponds to 5–10% of the remaining ice volume (Zemp et al. in press). The acceleration of the glacier shrinkage after 1985 indicates transition towards a rapid down-wasting rather than a dynamic glacier response to a changed climate (cf. Paul et al. 2004).

The general glacier retreat since 1850 corresponds well with the observed warming trend in this period (e.g. Oerlemans 1994, Maisch et al. 2000, Oerlemans 2001 p. 110–111, Zemp et al. subm.). However, the onset of the Alpine glacier retreat after 1850 might have been triggered by a negative winter precipitation anomaly (relating to the mean of 1901–2000) during the second half of the 19th century (Wanner et al. 2005). The intermittent periods of glacier advances in the 1890's, 1920's and 1970–1980's can be explained by earlier wetter and cooler periods, with reduced sunshine duration and increased winter precipitation (Patzelt 1987, Schöner et al. 2000, Laternser and Schneebeli 2003). Schöner et al. (2000) concluded from the investigation of a homogenised climate dataset and measured mass balance data from the Austrian part of the eastern Alps, that the more positive mass balance periods show a high correlation with winter accumulation and a lower correlation with summer temperature, while more negative mass balance periods correlate extremely well with summer temperature and show no correlation to winter accumulation. In addition they find that the positive mass balance period between 1960 and 1980 was characterised by negative winter North Atlantic Oscillation (NAO) index values, which caused an increase of the meridional circulation mode and a more intense north-westerly to northerly precipitation regime (cf. Wanner et al. 2005). The observed trend of increasingly negative mass balances since 1980 is consistent with accelerated global warming and correspondingly enhanced energy flux towards the earth's surface (FoG 2005).

Representativity of the SGI2000 and the fluctuation series

When analysing national inventories or individual fluctuation series, the question of representativity often arises. Is the investigated sub-sample and respectively, the surveyed glaciers, representative of the entire glacierisation? In this section we discuss the issue of representativity of the SGI2000, of the Alpine front variation series and of the Alpine mass balance series for the entire Alpine glacierisation.

The comparison of the area characteristics of the 1850- and 2000-sub-samples of the SGI2000 (on which the extrapolation of the Alpine areas of 1850 and 2000 are based) with the complete Swiss inventory in the WGI (Table 1, CH WGI) shows that the distributions of the area classes are similar. Nevertheless, small glaciers (<0.1 km^2) are under-represented, and glaciers in the north-eastern part of Switzerland are poorly represented (Paul et al. 2004). However, the SGI2000 sub-samples for 1850 and 2000 include 86% of the Swiss glaciers (WGI) covering 88% of the total area and 51% of the Swiss glaciers (WGI) covering 87% of the total area, respectively. Thus, the SGI2000 can be considered a representative sub-sample of the Swiss glacierisation (Table 1, WGI CH), which is highly similar to the glacierisation of the other Alpine countries (Table 1, AT, FR and IT). The ice coverage of the European countries is equally distributed with respect to the number of glaciers in each area class, with the largest glaciers being over-represented in the area distribution. Therefore, the different area classes were considered when the extrapolation was applied to all Alpine glaciers (Table 2). The large relative area change of the smaller glaciers leads to a more pronounced area change in the entire Alps than in Switzerland, which is due to many large glaciers of the Swiss Alps – when compared to the average size of glaciers throughout the Alps (assuming a uniform climate change over the Alpine region).

Front variations are mainly measured at mid-size and large glaciers while the glaciers smaller than 0.5 km^2 are under-represented (Table 1: column Alps, FoG, FV). This is obvious because glacierets and névés (perennial snow banks) are often not suitable for these kind of measurements in terms of accessibility and the low dynamic reaction, whereas mountain and valley glaciers are easier to reach, their front variations are much more pronounced and are related on glacier dynamics. Front variations series with more than nine measurement years exist for about 10% of all Alpine glaciers which cover more than 50% of the total glacier area. The dynamic response to climatic forcing of glaciers with variable geometry results in striking differences in the recorded curves, reflecting the considerable effects of size-dependent filtering, smoothing and enhancing of the delayed tongue response with respect to the undelayed input (mass balance) signal (Oerlemans 2001). Dynamic response time depends mainly on glacier length, slope and mass balance gradient (Johanneson et al. 1989, Haeberli and Hoelzle 1995). As a consequence, large valley glaciers with a dynamic response time of several decades show the secular climate trend while smaller mountain glaciers show marked periods of intermittent advances and retreats on a decadal scale. Smallest, somewhat static, low-shear stress glaciers (cirque glaciers) have altitude ranges that are comparable to or smaller than the inter-annual variability of equilibrium line altitude and hence, in general, reflect yearly changes in mass balance without any delay (Hoelzle et al. 2003).

Mass balance measurements are labour-intensive and are thus, only available from 25 glaciers, mainly with sizes from 0.5 to 10 km^2, covering only 3% of the glacier area (Table 1: column Alps, FoG, MB). In spite of their small number, they are geographically well distributed over the entire Alps (Figure 1). Mass balance is the direct and undelayed response signal to annual atmospheric conditions. It documents degrees of imbalance between glaciers and climate due to the delay in dynamic response caused by the characteristics of ice flow (deformation and sliding). Over long time intervals they indicate trends of climatic forcing. With constant climatic conditions (no forcing), balances would tend towards zero. Long-term non-zero balances are, therefore, an expression of ongoing climate change and continued forcing (GMBB 2005). Summer and winter balance even provide intra-annual climate information and should therefore be surveyed on all mass balance glaciers (Dyurgerov and Meier 1999, Vincent 2002). In general, Alpine fluctuation series are well distributed over the Alps and represent the range of area classes quite good. In view of the large contribution of glaciers smaller than 1 km^2 to glacier shrinkage in the past and the prediction towards ongoing global warming (e.g. Schär et al. 2004, Beniston 2005), future work should include studies on the influence of atmospheric warming on small glaciers and on current down-wasting processes (see also Paul et al. 2004). However, climatic sensitivity of glaciers not only depends on glacier size, but also on their sensitivity to regional climate variability versus local topographic effects, which potentially complicates the extraction of a regional or global climate signal from glacier fluctuations (Kuhn et al. 1985, Vincent et al. 2004). Mass balance and ice flow models calibrated with available fluctuation data are needed to quantify these effects (Oerlemans 1998, Oerlemans 2001, Paul et al. this issue).

6 Conclusions

In the European Alps the growth of the glacier monitoring network over time has resulted in an unprecedented glacier dataset with excellent spatial and temporal coverage. The WGMS compiled information on spatial glacier distribution from approximately 5,150 Alpine glaciers and fluctuation series (front variation and mass balance) from more than 670 of these glaciers. National inventories are able to provide complete Alpine coverage for the 1970's, when the glaciers covered an area of 2,909 km^2. This inventory, together with the SGI2000, is used to extrapolate Alpine glacier covered areas in 1850 and 2000 which amount to about 4,470 km^2 and 2,270 km^2, respectively. This corresponds to an overall glacier area loss from 1850 up until the 1970's of 35% and almost 50% until 2000.

Annual mass balance and front variation series provide a better time resolution of glacier fluctuations over the past 150 years than the inventories. During the general retreat, intermittent periods of glacier advances in the 1890's, 1920's and 1970–1980's can still be seen. Increasing mass loss, rapidly shrinking glaciers, disintegrating and spectacular tongue retreats are clear warnings of the atmospheric warming observed in the Alps during the last 150 years and the acceleration observed over the past two decades.

While inventory data contain information on spatial glacier distribution at certain times, fluctuation series provide temporal information at specific locations. Continuity and representativity of fluctuation series are thus most relevant for the planning of glacier monitoring. Furthermore, modelling should be enhanced and integrated into monitoring strategies. It is of major importance to continue with long-term fluctuation measurements and to extend the series back in time with reconstructions of former glacier geometries. Additionally, it is necessary to integrate glacier monitoring as well as reconstruction activities into the global framework of the GLIMS project (www.glims.org) and the WGMS (www.wgms.ch).

Acknowledgements

We are indebted to the numerous people who have provided the WGMS with data for many years! In particular, sincere thanks are therefore given to all the national correspondents of the WGMS as well as to G. Rossi and W. Schöner for the great collaboration during the revision of the Italian and the Austrian datasets. Special thanks go to S. Baumann, M. Dischl, A. Hinterberger, J. Meilwes, W. Peschke, C. Rothenbühler and A. Stolz for their contribution to the control, correction and formatting of the data and to I. Woodhatch for editing the English. We gratefully acknowledge the constructive comments of G. Pederson, an anonymous referee and the Scientific Editors. Last but not least we thank R. Frauenfelder and the other colleagues from the WGMS for their daily effort and the great team work. The present study is mainly funded by the ALP-IMP project within the EC-programme 'Environment/Global Change' (BBW. No. 01.0498-2) and the Department of Geography of the University of Zurich.

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Paper II

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The application of glacier inventory data for estimating past climate-change effects on mountain glaciers: a comparison between the European Alps and the Southern Alps of New Zealand

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Abstract

This study uses the database from national glacier inventories in the European Alps and the Southern Alps of New Zealand (hereinafter called the New Zealand Alps), which contain for the time of the mid-1970s a total of 5,154 and 3,132 perennial surface ice bodies, covering 2,909 km^2 and 1,139 km^2 respectively, and applies to the mid-1970s. Only 1,763 (35%) for the European Alps and 702 (22%) for the New Zealand Alps, of these are ice bodies larger than 0.2 km^2, covering 2,533 km^2 (88%) and 979 km^2 (86%) of the total surface area, respectively containing useful information on surface area, total length, and maximum and minimum altitude. A parameterisation scheme using these four variables to estimate specific mean mass balance and glacier volumes in the mid-1970s and in the '1850 extent' applied to the samples with surface areas greater than 0.2 km^2, yielded a total volume of 126 km^3 for the European Alps and 67 km^3 for the Southern Alps of New Zealand. The calculated area change since the '1850 extent' is -49% for the New Zealand Alps and -35% for the European Alps, with a corresponding volume loss of -61% and -48%, respectively. From cumulative measured length change data an average mass balance for the investigated period could be determined at -0.33 m water equivalent (w.e.) per year for the European Alps and -1.25 m w.e. for the 'wet' and -0.54 m w.e. per year for the 'dry' glaciers of the New Zealand Alps. However, there is some uncertainty in several unknown factors, such as the values used in the parameterisation scheme of mass balance gradients, which, in New Zealand vary between 5 and 25 mm m^{-1}.

Keywords: glacier fluctuations, glacier length changes, glacier mass changes, climate change, reconstruction, volume change, area change

1 Introduction

The last report of the Intergovernmental Panel on Climate Change (IPCC, 2001) stated that glaciers are the best natural indicators of climate. Hence, glacier changes are observed world wide within the framework of the GTN-G (Global Terrestrial Network on Glaciers) of the Global Climate Observing Systems (GCOS/GTOS, 1997a; GCOS/GTOS, 2004; Haeberli et al., 2000; Haeberli and Dedieu, 2004; Haeberli et al., 2002; WMO, 1997). The GTN-G is led by the World Glacier Monitoring Service (WGMS, http://www.wgms.ch), the successor to the 1894 founded international glacier commission (Forel, 1895). WGMS collects data based on the Global Hierarchical Observing Strategy (GHOST), which consists of five Tiers (GCOS/GTOS, 1997b; IUGG(CCS)/UNEP/UNESCO, 2005). An extensive amount of data on topographic glacier parameters, based on Tier 5 of the GHOST, has been built up in past regional glacier inventories (IAHS(ICSI)/UNEP/UNESCO, 1989), which are currently updated by modern remote sensing methods within the Global Land Ice Measurements from Space (GLIMS, http://www.glims.org) project (Bishop et al., 2004; Kieffer and others, 2000). Repetition of the glacier inventories should be undertaken at periods compatible with the characteristic dynamic response times of mountain glaciers (a few decades). However, current glacier down-wasting as observed in several mountain areas will probably require more updates of inventories in shorter time intervals (Paul et al., in press-a).

The glacier inventories help with assessing the problem of representivity of continuous measurements within different mountain areas, which can only be carried out on a few selected glaciers. A climate signal extracted from one single glacier is often not very representative for a whole mountain range. The understanding of global effects of climate change can only be achieved by comparing the long-term behaviour of glaciers within different mountain ranges. The European Alps were analysed in detail by Haeberli and Hoelzle (1995) and a reanalysis was recently done by Zemp et al. (in press-c). This data is now ready to compare with other mountain ranges. The Southern Alps of New Zealand (hereinafter called the New Zealand Alps) are particularly interesting for such a comparison, because they are situated at a similar latitude in the Southern Hemisphere as the European Alps in the Northern Hemisphere. However, both mountain ranges have quite different climatic conditions. The method used to assess climate change effects on glacier change in both regions are based on simple dynamic considerations and steady state conditions, which could be approximated for 1850 and the mid 1970s. Both inventories show a very high accuracy and are perfectly suited to apply the method described in detail by Haeberli and Hoelzle (1995). Based on this application the New Zealand inventory (Chinn, 1991; 2001) is compared with the European Alps in this study (Fig. 1). Both inventories are included in the WGMS database stored at WGMS in Zurich, Switzerland and at the National Snow and Ice Data Center (NSIDC, http://nsidc.org) in Boulder, Colorado, USA (Hoelzle and Trindler, 1998).

In this paper we show: (a) an application of an existing glacier parameterisation scheme to two glacier inventories in two different mountain ranges to compare characteristic variables like balance at the tongue, ice thickness, etc., (b) a method of mean specific mass balance and volume reconstruction based on observed equilibrium line altitude (ELA) changes, (c) a reconstruction of mean mass balance change by using glacier length change measurements and (d) a comparison of the different approaches within the different mountain ranges.

2 Methods

The parameterisation scheme developed by Haeberli and Hoelzle (1995) provides the possibility of analyzing large amounts of topographic glacier inventory data stored in the current databases at WGMS and NSIDC and will probably be archived in future databases at the NSIDC within the framework of the GLIMS project (Raup et al., 2003; in press). Detailed information about the full parameterisation scheme can be found in Haeberli and Hoelzle (1995). Therefore, we discuss only the most important parameterisations applied in this paper, based on theoretical concepts of Jòhannesson et al. (1989) and Nye (1960). We use only four measured input variables from the inventories, namely maximum altitude (H_{max}), minimum altitude (H_{min}), length (L_0) and total surface area (F). All other variables are calculated or taken from measurements (see Table 1). Where specific equations are not given, the corresponding references are cited.

This approach considers the step changes after full dynamic response and new equilibrium of the glacier has been achieved, when mass balance disturbance Δb leads to a corresponding glacier length change ΔL that depends on the original length L_0 and the average annual mass balance (ablation) at the glacier terminus b_t. The term b_t is calculated as $b_t = \delta b/\delta h \cdot (H_{mean} - H_{min})$, where H_{mean} is determined by $(H_{max} + H_{min}) / 2$:

$$\Delta b = b_t \cdot \Delta L / L_0 \qquad (1)$$

Glacier thickness (h) is determined according to equation (2) (Paterson, 1994) where α is the slope, τ the basal shear stress, ρ the density of ice and g the acceleration due to gravity (see Table 1 for used values of τ, ρ and g).

$$h = \tau / [\, \rho \cdot g \cdot \sin(\alpha)] \qquad (2)$$

The dynamic response time t_{resp} is calculated after Jòhannesson et al. (1989), where h_{max} is a characteristic ice thickness, usually taken at the equilibrium line where ice depths are near maximum. h_{max} is calculated as $2.5 \cdot h$, as estimated from known ice thickness measurements on various alpine glaciers world-wide (Bauder et al., 2003; March, 2000).

$$t_{resp} = h_{max} / b_t \qquad (3)$$

Assuming a linear change of the mass balance from b to zero during the dynamic response, the average mass balance $<b>$ can be calculated according to equation (4). $<b>$ values are annual ice thickness change (meters of water equivalent (w.e.) per year) averaged over the entire glacier surface, which can be directly compared with values measured in the field. Altough the method is quite simple, the results compare very well with long-term observations (Hoelzle et al., 2003). The factor n_{resp} denotes the count of possible response times for each glacier within the considered time period.

$$<b> = \Delta b / 2 \cdot n_{resp} \qquad (4)$$

The reaction time t_{react} is calculated, based on the kinematic wave velocity (Nye, 1965; Paterson, 1994) between the onset of the mass balance change and the first reaction at the glacier terminus from:

$$t_{react} = L_a / c \qquad (5)$$

where L_a is the length of the ablation area and c is taken as $4 \cdot u_{s,a,}$ (surface velocity in the ablation area) which corresponds quite well to observations (Müller, 1988).

3 Results

3.1 Glacierization in the 1970s

The inventories of the New Zealand Alps and the European Alps are well suited for a comparative study, because both have a high level of accuracy and both were compiled in the 1970s. The data bases from the national inventories used in this study contain a total of 5,154 perennial surface ice bodies in the European Alps and 3,132 for the Mountains of New Zealand. They were compiled for the early 1970s for the European Alps and 1978 for the Southern Alps of New Zealand. Only 1,763 (35%) for the European Alps and 702 (22%) for the Southern Alps of New Zealand of these total numbers are ice bodies larger than 0.2 km^2 having useful information available about surface area, total length, maximum and minimum altitude. All calculations presented here were performed with these two sub-samples. This is justified by the fact that by far the bulk of the ice is contained within the large glaciers, such as Aletsch glacier, which stores around 11% of the total ice mass within the European Alps and in New Zealand, nearly 50% is contained in the five largest glaciers. In addition, the applied parameterisation is better suited to larger ice bodies, because of their more distinct dynamical behaviour (Leysinger Vieli and Gudmundsson, 2004).

The measured surface area of the 1,763 glaciers in the European Alps and 702 glaciers in the Southern Alps of New Zealand > 0.2 km^2 are 2,533 km^2 and 979 km^2 corresponding to 88% and 86% of the total surface area, respectively. The calculated total volume of these glaciers is based on the ice thickness, equation (2), multiplied with the measured area in the 1970s and equals 126 km^3 for the European Alps and 67 km^3 for the New Zealand Alps. These volumes correspond to a sea-level rise equivalent to 0.35 mm for the European Alps and about 0.18 mm for the New Zealand Alps. These small values point to the limited significance for sea level rise, but mainly to the vulnerability to climate effects of glaciers in mountain areas with predominantly small glaciers (Barnett et al. 2005). Mountain ranges with such small glaciers could now be deglaciated within a few decades, whereas larger glaciers with high elevation ranges will persist somewhat longer before complete deglaciation. Small glaciers, especially in densely populated areas like the European Alps, have a strong impact on natural hazards, energy production, irrigations and/or tourism (Haeberli and Burn, 2002; Kääb et al., 2005a; Kääb et al., 2005b). The volume calculated for the Southern Alps of New Zealand (67 km^3) is somewhat higher than the value determined by Chinn (2001) and Heydenrych et al. (2002) of 53.3 km^3 and closer to a value determined earlier (63 km^3) by Anderton (1973). The mean maximum glacier elevation in the Southern Alps of New Zealand is 2,236 m a.s.l. ±289 m, and for the European Alps a value of 3,271 m a.s.l. ±322 m was found (Fig. 2a). Mean glacier elevations, which roughly equate to the equilibrium line altitude (ELA, here assumed as the steady-state ELA) is, for the European Alps, 2,945 m a.s.l. ±214 m and for New Zealand Alps 1,904 m a.s.l. ±220 m

(Fig 2b). The minimum elevation of the glaciers in the European Alps is 2,620 m a.s.l. ±264 m and for the New Zealand Alps a value of 1,545 m a.s.l. ±308 m was found (Fig 2c). The overall slopes of the glaciers are steeper in New Zealand (28.7°) than in Europe (24.2°) (Fig. 2d). The calculated average mass balance at the tongue (b_t) is, in the New Zealand Alps, around twice the ablation found in the European Alps. A maximum value was calculated as 23.9 m a^{-1} (Fox glacier) for the New Zealand Alps and 13.5 m a^{-1} (Bossons glacier) for the European Alps (Fig. 3). Calculated response times (the time taken to reach equilibrium after a 'step' climate change) after equation (3) have a mean value of 11.6 years in the New Zealand Alps and 37.4 years in the European Alps (Fig. 4). As an example, the calculated response time for Franz Joseph glacier is around 20 years, which corresponds well to values found with a numerical model by Oerlemans (1997). The reaction time of Franz Joseph glacier is around seven years as estimated in this study (Table 3b). This value corresponds very well with the values found in other studies (Chinn, 1996; Hooker, 1995; Hooker and Fitzharris, 1999; Woo and Fitzharris, 1992). In contrast, the response time of Aletsch Glacier in the European Alps is around 80 years (Table 3b).

Even though the sub-samples were selected for larger glaciers, these glaciers mainly have surface areas smaller than 10 km^2, lengths shorter than 5 km and overall slopes steeper than 10°. This means that the sample of presently existing alpine glaciers is dominated by small and steep mountain glaciers with average thicknesses of a few tens of meters (see Fig. 5).

3.2 Reconstruction of the mean specific mass balances

The reconstruction of the mean specific mass balances since the '1850 extent' is valuable because it is directly comparable to present day measurements and therefore to current trends. The precise year of the maximum glacier extension in the Little Ice Age (LIA) period is often difficult to determine (Grove, 1988). In the European Alps, the maximum glacier extension differs considerably from glacier to glacier (e.g. Aletsch 1859–1860, Gorner 1859–1865 and Unterer Grindelwald 1820–1822, see Holzhauser et al., 2005) and for New Zealand, recent datings (Winkler, 2004) and studies at Franz Joseph glacier (Anderson, 2003) have shown that maximum LIA glacier extension was at the end of the 18th century, but that there was only minor glacier retreat to the end of the 19th century. Therefore, here the time of LIA maximum is arbitrarily set at 1850 AD although there are numerous cases where different dates for the maximum extents have been found.

Kuhn (1989; 1993) calculated that a temperature change of +1°C would increase the ELA by 170 m with an accuracy of ±50 m per 1°C. Chinn (1996) reported a possible shift in ELA of up to 200 m for the Southern Alps of New Zealand, which corresponds to an air temperature increase of about 1.2°C, which is in good agreement with a measured temperature change suggested by Salinger (1979) of around 1°C for New Zealand. If the ELA-change is known, the corresponding change in mass (Δb) can be calculated together using the mass balance gradient ($\delta b/\delta h$) after equation (6).

$$\Delta b = \delta b/\delta h \cdot \delta ELA/\delta T_{air} \cdot \Delta T_{air} \qquad (6)$$

where $\delta ELA/\delta T_{air}$ describes the vertical shift of ELA per 1°C, and integrates the change of all climate parameters, i.e. radiation, humidity, accumulation and air temperature, as well as feedback effects for any temperate glaciers (Kuhn, 1993).

Mass balance gradients play a very important role in the parameterisation scheme. It is therefore fundamental to select realistic gradients for each area. For the European Alps a mass balance gradient of 7.5 mm m^{-1} was chosen for the whole sample. In contrast, the New Zealand Alps are characterized by much stronger precipitation gradients in comparison to the European Alps (Chinn et al., 2005a; Fitzharris et al., 1997), with the associated large range of mass balance gradients. Although, several studies in the New Zealand Alps have shown that there are no large differences in response between glaciers and climate of the glaciers west and east of the Main Divide (Chinn, 1996; 1999 and Chinn et al., 2005a), mass balance gradients are completely different between glaciers west and east of the Main Divide (Chinn et al., 2005a). Therefore, the New Zealand sample was divided into two sub-samples, 'wet' to the west with a mass balance gradient of 15 mm m^{-1} and 'dry' to the east with a mass balance gradient of 5 mm m^{-1}. The 'wet' mass balance gradients were applied to three regions designated 'west', 'fiord' and 'east_wet' and the 'dry' mass balance gradients applied to 'east_dry' and 'north_dry' (Chinn et al., 2005a) (Fig. 1b). In the New Zealand Alps, an ELA shift of 200 m, will induce a mass balance change of 3.0 m for the 'wet' glaciers and 1.0 m for the 'dry' glaciers. The calculation of the $<b>$ value was done by taking into account each individual response time and multiples thereof (equation 4). The results of these calculations are presented in Fig. 6. The values range between -0.67 m w.e. (water equivalent) in the 'west' region and -0.54 m w.e. in the 'east_dry' region. No differentiation for the mass balance gradient was made for the European Alps, because there are no clear regional differences although, locally mass balance gradients can vary strongly.

3.3 Reconstruction of the 1850 area/volume and subsequent losses

In determining the volume change since the '1850 extent' and the original volume at 1850, the volume was calculated according to the parameterisation method described in Haeberli and Hoelzle (1995). Using the calculated mean specific mass balance for each area, the total mass loss could be estimated from the mean area of the 1970s and the 1850 area (F_{1850}), given in the following equation:

$$F_{1850} = 0.002 + 0.285 \cdot L_{1850} + 0.219 \cdot (L_{1850})^2 - 0.004 \cdot (L_{1850})^3 \qquad (7)$$

where L_{1850} is the estimated length at the end of the Little Ice Age, calculated from equation (1). The third-degree polynomial fit to the data is chosen to avoid negative area values for the smallest glaciers and to optimally reproduce the length/area-relation for large valley glaciers (Haeberli and Hoelzle, 1995). The so-calculated F_{1850} is around twice the area of 1978 for the New Zealand Alps, and around 1.5 times the area of the 1970s for the European Alps. However, this result not only neglects the area changes for the ice bodies < 0.2 km^2 in the current inventories, but also excludes all ice bodies, which had completely disappeared before the mid 1970s. The reconstructed area of Zemp et al. (in press-c), included both the small and vanished glaciers, and is therefore a value of 4,470 km^2. Their corresponding value of area loss from 1850 to 1970 was -35% and is equal to the calculated loss of this study. Calculated ice volumes are 126 km^3 for the 1970s in the European Alps and 67 km^3 for 1978 for the New Zealand Alps (see details in Table 2). The calculation is based on the thickness along the flowline, equation (2). Volume loss is determined from the calculated mean specific mass balances for each region and the mean area for the 1970s and 1850. Around 61% of the original volume has been lost in the New Zealand Alps and around 48% in the European Alps (Table 2 and Fig. 7, 8).

3.4 Reconstruction of the mean specific mass balance based on length change measurements

The next task was to simulate the 1850 glacial conditions and to check how realistic the proposed scheme is. Present length change measurements were taken from monitored glaciers in the European Alps and the New Zealand Alps (Chinn, 1996). From these length changes, the mean mass balance since 1850 was determined by using equation (1) in the reversed manner as used in the previous two subsections (Hoelzle et al., 2003). Only those glaciers were selected, which are not calving in lakes or are heavily debris covered, as these types of glaciers are frequently divorced from climate forcing by accelerated retreat (lakes) and thermal insulation (debris coverd). Many of the large glaciers in New Zealand are heavy debris covered and/or calving into proglacial lakes.

The time interval considered is 125 years for the European Alps and 128 years for the Southern Alps of New Zealand. A single step positive mass balance change of 1 meter w.e. per year was assumed for the entire time interval for the European Alps and the ‘dry’ glaciers for the New Zealand Alps. For the ‘wet’ glaciers in New Zealand a value of 3 meters was chosen. A comparison between calculated and geomorphologically reconstructed length changes for selected glaciers of the New Zealand ‘dry’ and ‘wet’ glacier samples show that the different sensitivities of long-term glacier length change as a response to a uniform mass balance forcing can be quite well reproduced (Table 3) and that the chosen mass balance forcing appears to slightly underestimate the real evolution, especially for glaciers in the ‘wet’ sample. Differences between measured and calculated overall length changes for individual glaciers can be considerable and are explained by the uncertainties in the parameterisation scheme applied, and by variable climate/mass balance conditions at each glacier. The mass balance forcing for each glacier can be corrected according to equation (8) to fit the measured length changes.

$$b^* = \Delta L_m / \Delta L_c \cdot \langle b \rangle_i / 1 \qquad (8)$$

where b^* is the corrected average mass balance, ΔL_m is the measured length change, ΔL_c is the calculated length change and $\langle b \rangle_i$ is the mean mass balance for the individual regions. These corrections from equation (8) to the mass balance forcing for each glacier to fit the measured length change gives an average annual mass balance (b^*) of -0.33 ±0.09 m w.e. per year for glaciers in the European Alps. In New Zealand, the average annual mass balance for the ‘wet’ glaciers is -1.25 ±0.9 m w.e. per year and for the ‘dry’ glaciers -0.54 ±0.5 m w.e. per year. The calculated mass balance change Δb for the ‘wet’ glaciers in the New Zealand Alps is around 5 m and for the ‘dry’ glaciers around 1 m. The latter value corresponds quite well with the value used in the parameterisation scheme, whereas the value for the ‘wet’ glaciers is, much higher than the value of 3 m used in the parameterisation. This suggests that the glaciers have reacted even more sensitively than assumed in our parameterisation and that it is possible that our calculated mass loss is at the lower boundary of the uncertainty range.

4 Discussion and conclusions

The calculations and estimations presented in this study are built on four very simple geometric parameters contained in detailed glacier inventories. This justifies the simplicity of the applied algorithms but also means that the uncertainties involved with the proposed procedure are considerable. Indeed, the large scatter in derived parameters such as mass balance at the tongue, response times etc. points to the fact that the applied parameterisation scheme is more useful for relatively large glaciers than for small ice bodies. The large glaciers dominate the overall mass changes and hence, make the estimates of corresponding changes probably quite realistic. This is clearly indicated by the parameterisation results of the large glaciers such as Aletsch or Franz Joseph, which show quite realistic computation results (Table 3b) in comparison to detailed numerical studies (Oerlemans, 1997). The model is not tuned in any kind to the European Alps, where the model has been applied for the first time. The only factor changed in the model is the mass balance gradient and the basal shear stress (see Table 1). The values used for these parameters in the study are all based on measurements. In any case, the striking sensitivity of glacierization in mountain areas to atmospheric warming trends clearly appears in both mountain regions, although some marked differences do exist. The calculations of the mean specific mass balance change for the investigated period show no large differences between maritime and the ‘dry’ glaciers within New Zealand (Fig. 6). However, the mean specific mass balances calculated for New Zealand are close to twice than the values determined for the European Alps and indeed for other mountain areas (Hoelzle et al., 2003). In addition, the calculations show that the relative mass loss seems to be considerably larger in the New Zealand Alps than in the European Alps and the comparison of the measured length changes suggests an even more pronounced difference. Therefore, the calculated mass loss given here for the Southern Alps of New Zealand is probably at the lower limit of the uncertainty range.

The ongoing glacier behaviour after the mid-1970s was sometimes quite different between the European Alps and the New Zealand Alps. After a period of glacier advance in the 1970s and early 1980s in the European Alps, the glaciers experienced a strong retreat during the following 20 years with an even more pronounced mass loss at the beginning of the 21st century (Paul et al., 2004). Today the glacier surface area in the European Alps is around 2,270 km^2 (Zemp et al., in press-c). In contrast, all glaciers in New Zealand have overall experienced a positive mass balance and some have advanced strongly during the 1990s. This is especially true for the ‘wet’ glaciers in New Zealand; the sample is clearly dominated by maritime, highly sensitive glaciers with corresponding high precipitation and therefore strong mass turn over. This period of advances was coming to an end at the beginning of the new century (Chinn et al., 2005b). The glaciers in the New Zealand Alps will probably react more sensitively to a future temperature increase than those in the European Alps. Not only because of their greater sensitivity, as expressed by the mass balance gradient, but also because of the generally low altitude of the ELA in the New Zealand Alps promoting a higher percentage of rain rather than snowfall in the future.

Due to increasing uncertainties and pronounced non-linearities such as changing response times with changing glacier size etc., calculations for scenarios with a trend to continuously accelerate climate and mass balance forcing beyond the early decades in the 21st century can be order-of-magnitude-estimates only. Annual mass losses of two to three meters per year such as observed in the year 2003 in the European Alps (Frauenfelder et al., 2005; IUGG(CCS)/UNEP/UNESCO/WMO, 2005; Zemp et al., in press-a) and which must be expected to continue into the future, IPCC scenarios of temperature increase will certainly reduce the surface area and volume of alpine glaciers to a few percent of the values estimated for the ‘1850 extent’ within the 21st century. With such a trend, only the largest and highest-reaching alpine glaciers could persist into the 22nd century. These glaciers will be affected by drastic changes in geometry as well and down-wasting rather than active retreat will be the dominant process of glacier reaction (Paul et al., 2004; Paul et al., in press-a). This implies that parameterisation schemes like the one presented here would no longer be applicable. Therefore, modern technologies like remote sensing have to be used to measure the accelerating glacier change and alpine wide mass balance (Machguth et al., in press; Paul et al., in press-b) and ELA models (Zemp et al., in press-b) need to be applied for the extrapolation of glacier down-wasting into the future.

This comparison between two different Alpine regions demonstrates the potential and limitations of the parameterisations of existing inventories. Their use lies especially in quantitatively inferring past average decadal to secular mean specific mass balances for unmeasured glaciers by analyzing cumulative length change from moraine mapping, satellite imagery, aerial photography and long-term observations (Haeberli and Holzhauser, 2003).

Acknowledgments

We would like to thank Ivan Woodhatch for his help in improving the English in a first version of this paper. The present study was carried out within the data analysis work for the World Glacier Monitoring Service with special grants from UNEP, a personal research grant from the Swiss National Science Foundation (PIOI2 – 111008/1) and as part of the EU Research Programme ALP-IMP (EU: EVK2-CT-2002-00148, BBW: 01.0498-2). The authors are very grateful to the constructive comments of Stephan Winkler, Per Holmlund and Christoph Schneider, which helped to improve the paper.

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Tables

Table 1: Parameterised values used in the calculations.

Parameters	EU Alps	NZ Alps 'wet'	NZ Alps 'dry'
τ [Pa]	$1.3 \cdot 10^5$	$1.8 \cdot 10^5$	$1.2 \cdot 10^5$
$\delta b/\delta h$ [mm m^{-1}]	15	5	7.5
ρ [kg m^{-3}]	900	900	900
g [m s^{-2}]	9.81	9.81	9.81
A [a^{-1} Pa^{-8}]	0.16	0.16	0.16
n	3	3	3

Table 2: Data used and calculated for all glaciers > 0.2 km^2

Regions	Area 1970s [km^2]	Area LIA [km^2]	Count of glaciers	<b> used [m a^{-1}]	Volume 1970s [km^3]	Volume LIA [km^3]	Volume loss [km^3]	Volume loss [%]
North_dry	0.69	3.81	2	-0.6	0.0079	0.16	-0.15	
East_wet	350.26	640.43	204	-0.61	32.37	66.40	-34.03	
East_dry	122.80	257.39	129	-0.53	5.367	16.71	-11.35	
West	464.20	951.67	301	-0.67	27.56	80.98	-53.43	
Fiord	40.80	78.36	63	-0.65	1.467	5.82	-4.36	
NZ Alps	978.75	1931.66	702		66.77	170.10	-103.33	-61
EU Alps	2544.38	3914.61	1763	-0.33	126	241.35	-115.35	-48

Table 3: Three selected glaciers from the parameterisation scheme: Fox, Franz Joseph and Aletsch. (a) shows the input parameters for the parametrisation scheme. (b) shows some selected calculated output parameters (u_d = deformation velocity, u_b = sliding velocity, u_s surface velocity are calculated after Haeberli and Hoelzle (1995)).

(a) Glaciers	H_{max} [m a.s.l.]	H_{mean} [m a.s.l.]	H_{min} [m a.s.l.]	Area 1970s [km^2]	Length 1970s [km]	Length change measured (ΔL_m) [km]
Fox	3500	1900	306	34.69	13.20	2.50
Franz Joseph	2955	1690	425	32.59	10.25	2.95
Aletsch	4140	2830	1520	86.76	24.70	2.50

(b) Glaciers	t_{react} [a]	t_{resp} [a]	α [°]	*Volume* [10^6 m^3]	u_d [m a^{-1}]	u_b [m a^{-1}]	u_s [m a^{-1}]	b_t [m]	h_{max} [m]	Length change calculated (ΔL_c) [km]	$\Delta L_m/\Delta L_c$	b^* [m]
Fox	6.0	16.7	13.6	2505.3	50.5	359.4	410	23.9	400.5	1.66	1.51	-1.13
Fr. Joseph	7.5	20.7	13.9	2310.9	49.6	207.6	257.2	19.0	392.4	1.62	1.82	-1.37
Aletsch	27.3	76.6	6.1	13719.1	54.4	115.1	169.5	9.8	752.7	2.50	1.00	-0.25

Figures

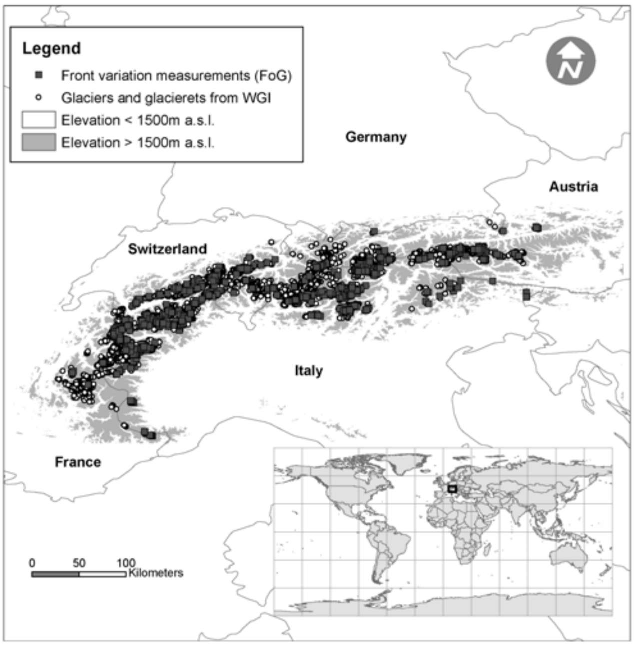

Figure 1a): Distribution of inventory glaciers (WGI = World Glacier Inventory) and of glaciers with length change measurements (FoG = Fluctuations of Glaciers) in the European Alps.

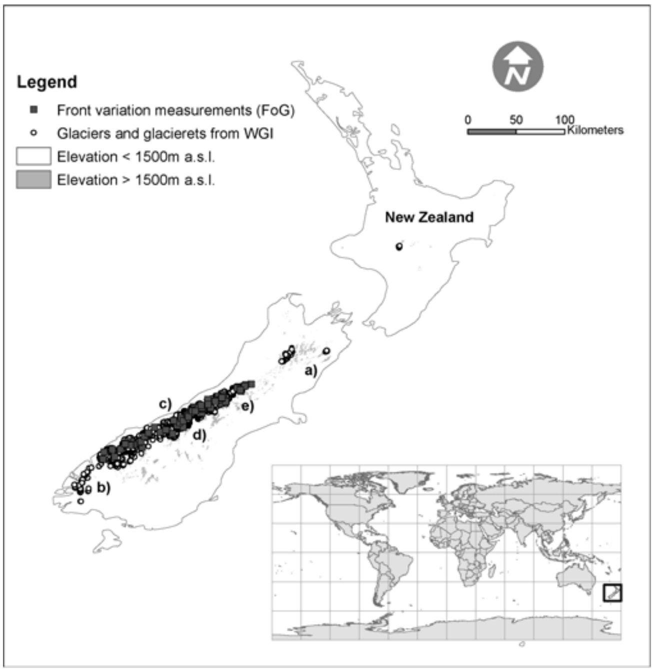

Figure 1b) Distribution of inventory glaciers (WGI) and of glaciers with length change measurements (FoG) in New Zealand. Only glaciers on the Southern Island of New Zealand were used in the analysis. The letters (a) to (g) are used to differentiate between 'wet' and 'dry' glaciers in the Southern Alps of New Zealand (a) = North_dry, (b) Fiord ('wet'), (c) West ('wet'), (d) East_wet and (e) East_dry). The digital elevation model used to draw the elevations is based on http://edc.usgs.gov/products/elevation/gtopo30/hydro/index.html.

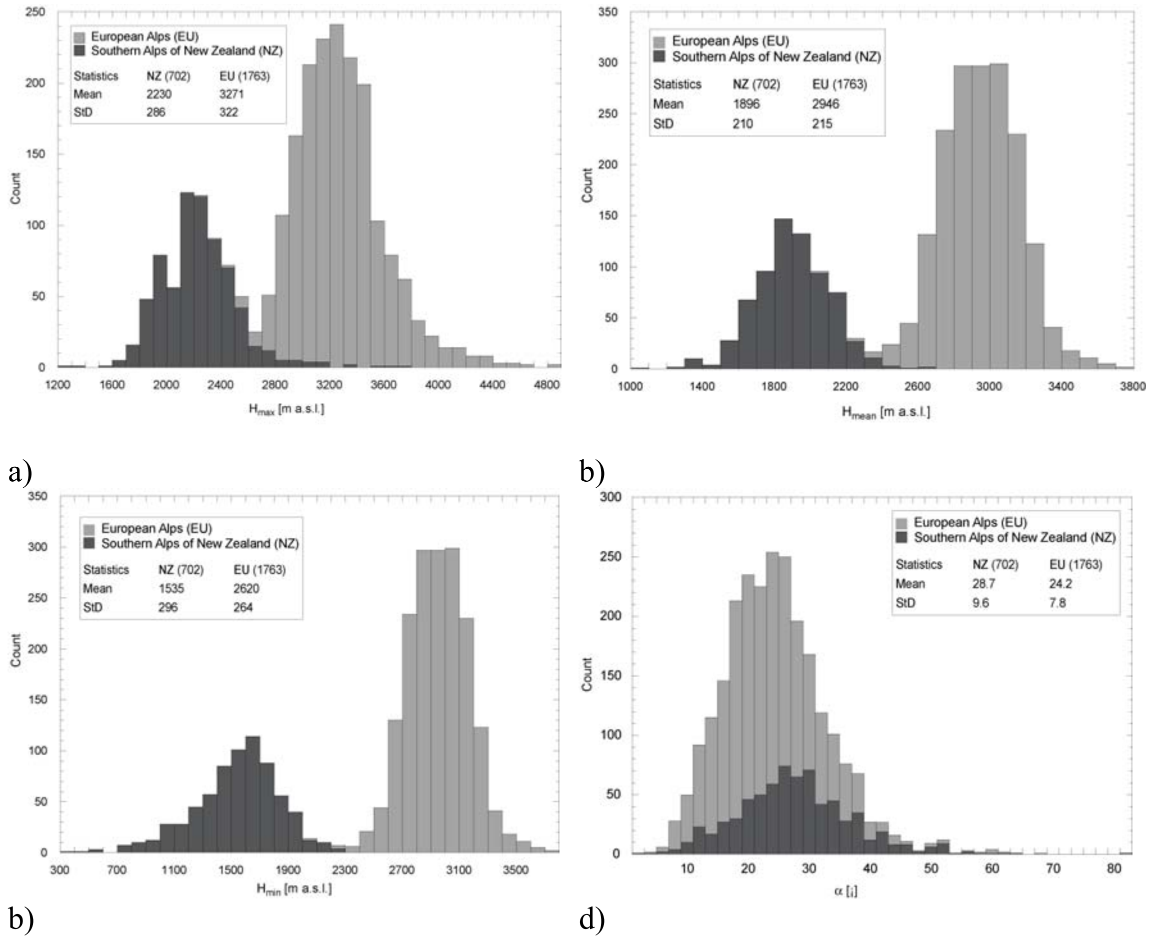

Figure 2: Distribution of: a) maximum (H_{max}), b) mean (H_{mean}), c) minimum (H_{min}) altitude and d) the slope (α) of inventory glaciers of the European Alps and the Southern Alps of New Zealand.

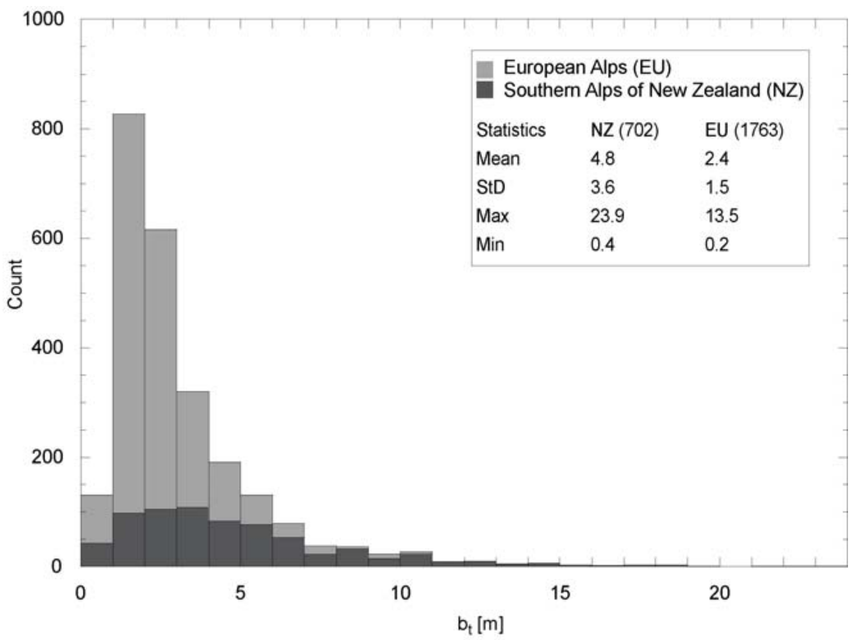

Figure 3: Calculated values of mass balance at the tongue (b_t) the inventory glaciers of the European Alps and the Southern Alps of New Zealand.

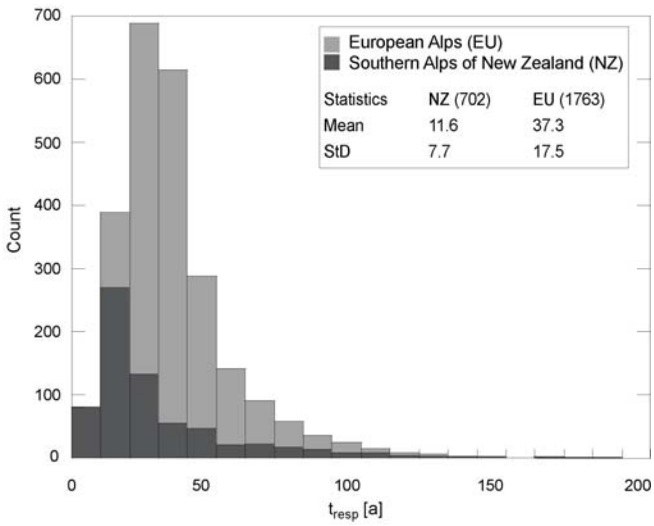

Figure 4: Calculated values of response times (t_{resp}) for the inventory glaciers of the European and Southern Alps of New Zealand.

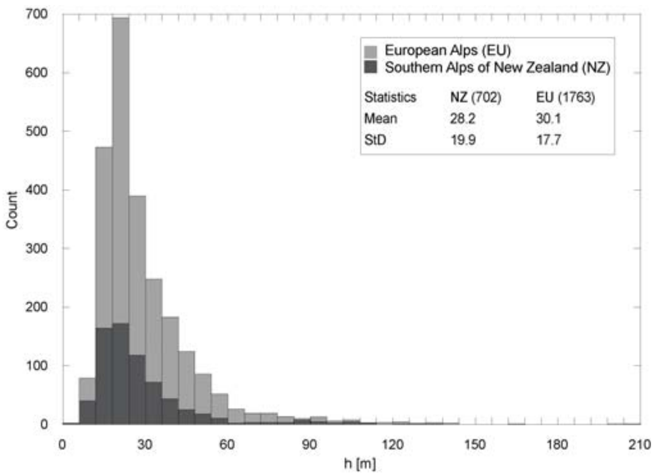

Figure 5: Calculated values for mean ice thickness (h) along the flowlines for the inventory glaciers of the European Alps and the Southern Alps of New Zealand.

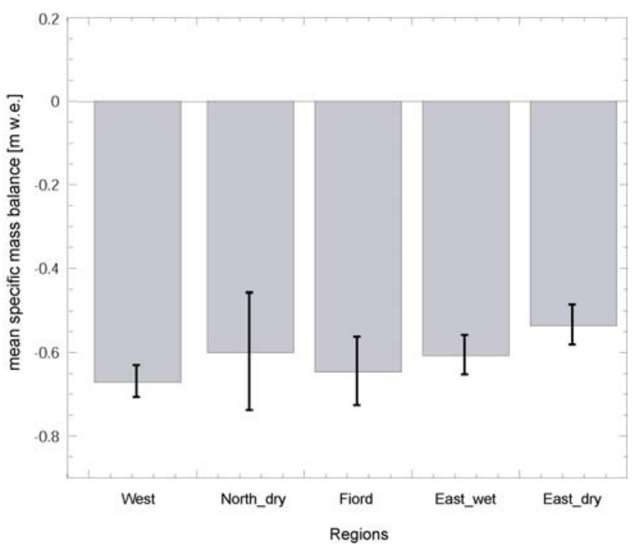

Figure 6: Calculated mean specific mass balance (<b>) for the time period between the mid 1970s and the '1850 extent' for the different regions (a) to (e) for the Southern Alps of New Zealand. The error bars are only a statistical value for the standard error.

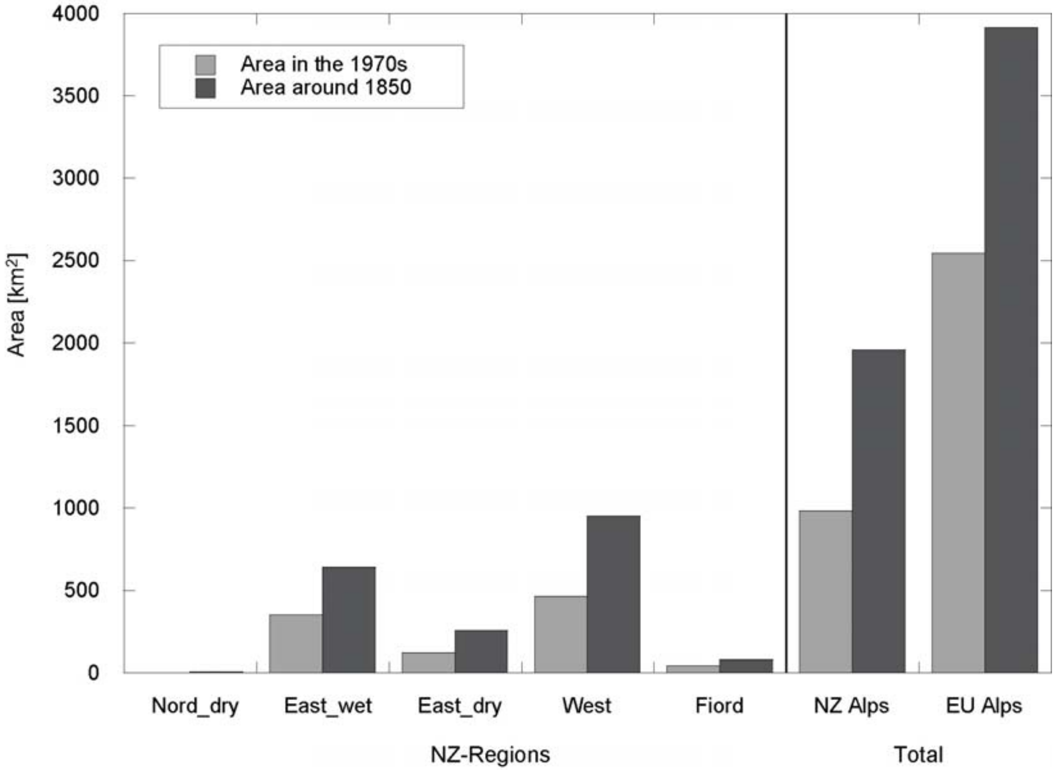

Figure 7: Calculated areas for the time of the mid 1970s and the '1850 extent' in the European Alps and the Soutern Alps of New Zealand.

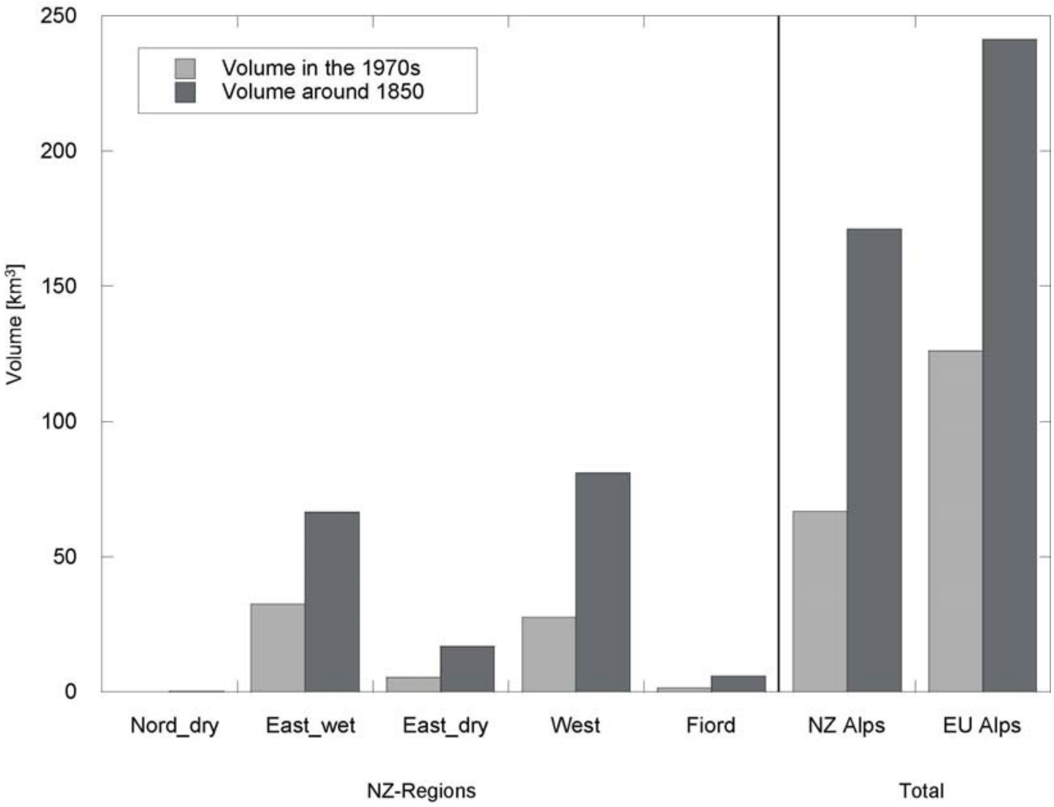

Figure 8: Calculated volumes for the time of the mid 1970s and the '1850 extent' in the European and Southern Alps of New Zealand.

Paper III

Worldwide glacier mass balance measurements: general trends and first results of the extraordinary year 2003 in Central Europe

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World Glacier Monitoring Service (WGMS), Zürich, Switzerland

Рассмотрены тренды баланса массы ледников мира, в частности, за 1980–2001 гг. и для чрезвычайно жаркого и сухого лета 2003 г. в Центральной Европе, которое привело к потере 5–10% объема ледников Альп.

Introduction

Worldwide collection of information about ongoing glacier changes was initiated in 1894 with the foundation of the International Glacier Commission at the Sixth International Geological Congress in Zürich, Switzerland. It was hoped that long-term glacier observations would give insight into processes of climatic change such as the formation of ice ages. In 1986 the World Glacier Monitoring Service (WGMS) started to maintain and continue the collection of information on ongoing glacier changes, when the two former ICSI services PSFG (Permanent Service on Fluctuations of Glaciers) and TTS/WGI (Temporal Technical Secretary/World Glacier Inventory) were combined [4].

Since its initiation, the goals of international glacier monitoring have evolved and multiplied. Today, the WGMS is integrated into global climate-related observation systems and collects standardized observations on changes in mass, volume, area and length of glaciers with time (glacier fluctuations), as well as statistical information on the distribution of perennial surface ice in space (glacier inventories). Thus, a valuable and increasingly important data basis on glacier changes has been built up over the past century [4].

International assessments such as the periodical reports by the Intergovernmental Panel on Climate Change (IPCC) or the GCOS/GTOS Plan for Terrestrial Climate-related Observation [1] define mountain glaciers as one of the best natural indicators of atmospheric warming with the highest reliability ranking. The Global Terrestrial Network for Glaciers (GTN-G) of the Global Terrestrial Observing System (GTOS), aims at combining (a) in-situ observations with remotely sensed data, (b) process understanding with global coverage and (c) traditional measurements with new technologies by using an integrated and multi-level strategy. This approach, the Global Hierarchical Observing Strategy (GHOST), uses observations in a system of Tiers. These Tiers include extensive glacier mass balance measurements within major climatic zones for improved process understanding and the calibration of numerical models (Tier 2) as well as the determination of regional glacier volume change within major mountain systems using cost-saving methodologies (Tier 3). A network of 60 glaciers representing Tiers 2 and 3 is already established. This data sample closely corresponds to the data compilation published so far by the WGMS with the bi-annual «Glacier Mass Balance Bulletin» [6].

The present contribution gives an overview on presently observed rates of change in worldwide mass balance of mountain glaciers, corresponding trends and regional peculiarities, such as the extremely hot and dry Central European summer of 2003.

Worldwide glacier mass balance observations 1980–2001

Glacier fluctuations result from changes in the mass and energy balance at the Earth's surface. In ablation and temperate firn areas, which predominate at lower latitudes/altitudes and in regions with humid climatic conditions, atmospheric warming causes mainly changes in the mass and geometry of glaciers [3]. Hereby, glacier mass balance is the direct undelayed signal of climatic change, whereas changes in glacier length primarily constitute an indirect, delayed and filtered but also enhanced signal (Fig. 1). Cumulative mass changes lead to ice thickness changes which, in turn, exert a positive feedback on mass balance and at the same time, influence the dynamic redistribution of mass by glacier flow [3].

As mentioned above a network of approximately 60 glaciers with long-term mass balance measurements provides information on presently observed rates of change in glacier mass, corresponding acceleration trends and regional distribution patterns (Fig. 2). Continuous mass balance records for the period 1980–2001 are now available for 30 glaciers for the period 1980–2000 and for 29 glaciers in the year 2000/2001 [6].

These values show that the mean for the period 1990–1999 (-373 mm w.e.) was twice as high as the corresponding mean of the previous decade of 1980–1989 (-188 mm w.e.), with even the year with the maximum mean (1993) yielding a slightly negative mean specific net balance (-9 mm w.e.).

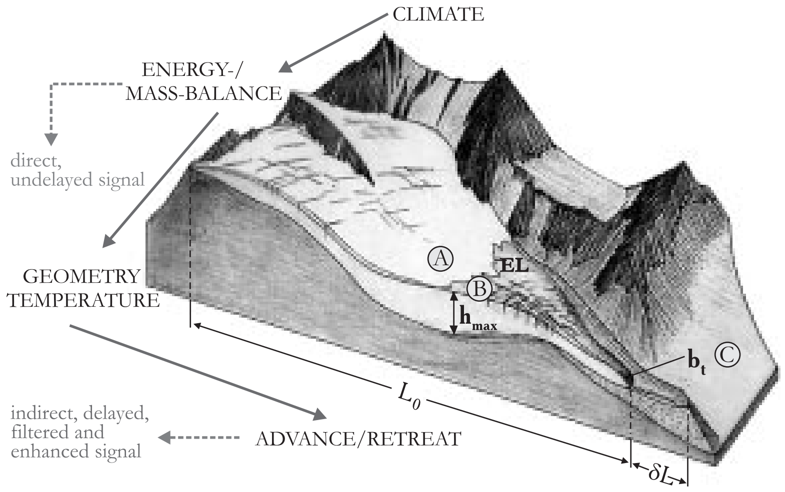

Fig. 1. Schematic plot illustrating the response of glaciers to climatic changes, with glacier mass balance as the direct undelayed signal, and changes in glacier length as an indirect, delayed and filtered but also enhanced signal. Figure modified after [3]

Рис. 1. Схема, иллюстрирующая реакцию ледников на изменения климата, где баланс массы служит прямым немедленным сигналом, а изменения длины ледника — опосредованным, отфильтрованным, хотя и усиленным сигналом, проявляющимся с задержкой. Рисунок с дополнениями по [3]

When considering these values, it has to be taken into account that the mean of all glaciers included in the observed sample is strongly influenced by the large proportion of Alpine and Scandinavian glaciers. Therefore, a mean value is calculated using only one single value (in places averaged) for each of the 9 mountain ranges concerned (Cascades, Alaska, Andes, Svalbard, Scandinavia, Alps, Altai, Caucasus, Tien Shan). Mean specific net balance for the 9 mountain regions during the period 1990–1999 amounts to -482 mm w.e., a value three times higher than the corresponding value for the previous decade from 1980 to1989 (-149 mm w.e.). For the time period from 1980 to 2001 mean specific net balance in these mountain regions averaged roughly -0.3 m w.e. with 20 negative and two positive balance years during these 22 years. The range of extremes observed at individual glaciers during one measurement year is roughly one order of magnitude higher than the mean value of the sample (Fig. 3, a). The significance of the recorded signal, on the other hand, increases with mass balance values cumulated over extended time periods (Fig. 3, b).

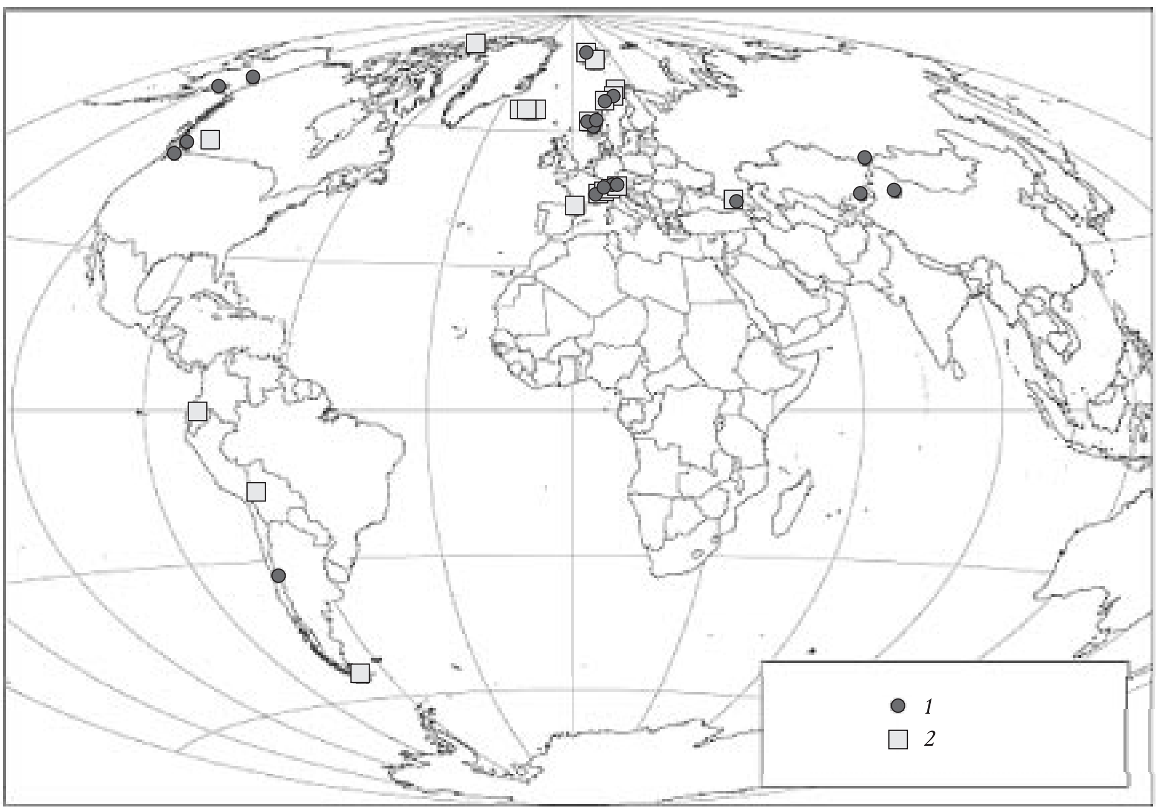

Fig. 2. Current worldwide glacier mass balance monitoring: 1 — glaciers with continuous records from 1980–2001, 2 — glaciers with interrupted or shorter series. Continents and country boundaries provided by ESRI

Рис. 2. Современное состояние мониторинга баланса массы ледников мира: 1 — ледники с непрерывными рядами за 1980–2001 гг., 2 — ледники с пропусками в рядах или более короткими рядами наблюдений. Границы континентов и стран предоставлены ESRI

Table 1

Mean specific net mass balance of 30 glaciers worldwide for the decades 1980–1989 and 1990–1999, and for the period 1980–2001; values in mm w.e.

	Mean 1980–1989	Mean 1990–1999	Mean 1980–2001
Minimum mean	-516	-712	-712
Maximum mean	+112	-9	+112
Mean	-188	-373	-275
Std. Deviation	±243	±234	±245

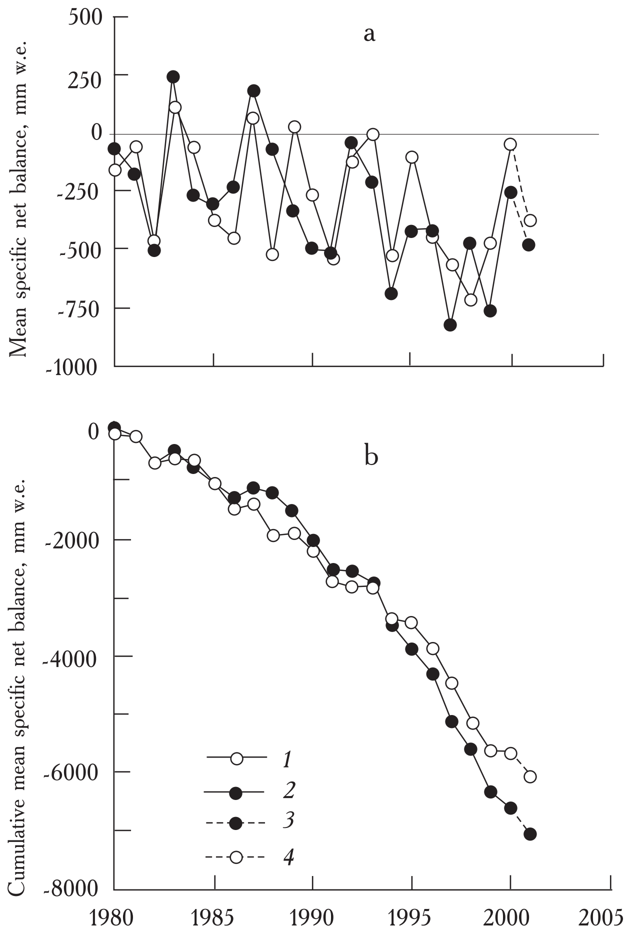

Fig. 3. Evolution with time of the mass balance of 30 glaciers in 9 mountain ranges worldwide (1980–2000), respectively for 29 glaciers in 8 mountain ranges (2001); (a) mean specific net balance, (b) cumulative mean specific net balance [6]: 1, 2 — mean of 30 glaciers and of 9 mountain ranges, 3, 4 — mean of 29 glaciers and 8 mountain ranges, correspondingly

Рис. 3. Изменения во времени баланса массы 30 ледников в 9 горных районах мира (1980–2000) и соответственно для 29 ледников в 8 горных районах (2001); средний удельный баланс (a), кумулятивный средний удельный баланс (b) [6]: 1, 2 — среднее для 30 ледников и 9 хребтов, 3, 4 — среднее для 29 ледников и 8 хребтов, соответственно

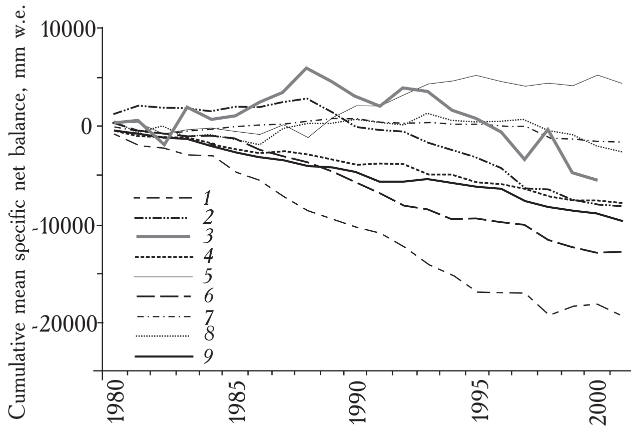

Fig. 4. Cumulative mean specific net balance for 9 different mountain regions for the period 1980–2001: 1 — Cascade Mountains (2 glaciers), 2 — Alaska (2 glaciers), 3 — Andes (1 glacier, 1980–2000), 4 — Svalbard (2 glaciers), 5 — Scandinavian peninsula (8 glaciers), 6 — Alps (9 glaciers), 7 — Altai (3 glaciers), 8 — Caucasus (1 glacier), 9 — Tien Shan (2 glaciers)

Рис. 4. Кумулятивный средний удельный баланс для 9 разных горных районов мира за 1980–2001 гг.: 1 — Каскадные горы (2 ледника), 2 — Аляска (2 ледника), 3 — Анды (1 ледник, 1980–2000 гг.), 4 — Шпицберген (2 ледника), 5 — Скандинавия (8 ледников), 6 — Альпы (9 ледников), 7 — Алтай (3 ледника), 8 — Кавказ (1 ледник), 9 — Тянь-Шань (2 ледника)

With the exception of maritime regions (e.g. Alaska, Scandinavia), cumulative trends for the individual mountain ranges (Fig. 4) show continued mass loss since the beginning of the period observed (1980–2001). A special case is the measurements series from the Andes, here, only one glacier is considered (Echaurren Norte, Chile) which is strongly influenced by the ENSO-phenomena. Therefore, this series has to be interpreted carefully in view of possible climatically induced trends (Fig. 4).

In summary, mean annual loss in ice thickness of mountain glaciers during 1980–2001 amounts to approximately -0.3 m w.e. per year, resulting in a total thickness reduction of approximately 6 to 7 m of ice since 1980 (see Fig. 3, b).

Extraordinary situation in Central Europe during 2003

Weather. In the following section information comes, if not stated differently, from reports of the Swiss Federal Institute for Snow and Avalanche Research [19] and the Swiss National Weather Service [12, 13]. Location of the place names mentioned in the text are shown in Fig. 5.

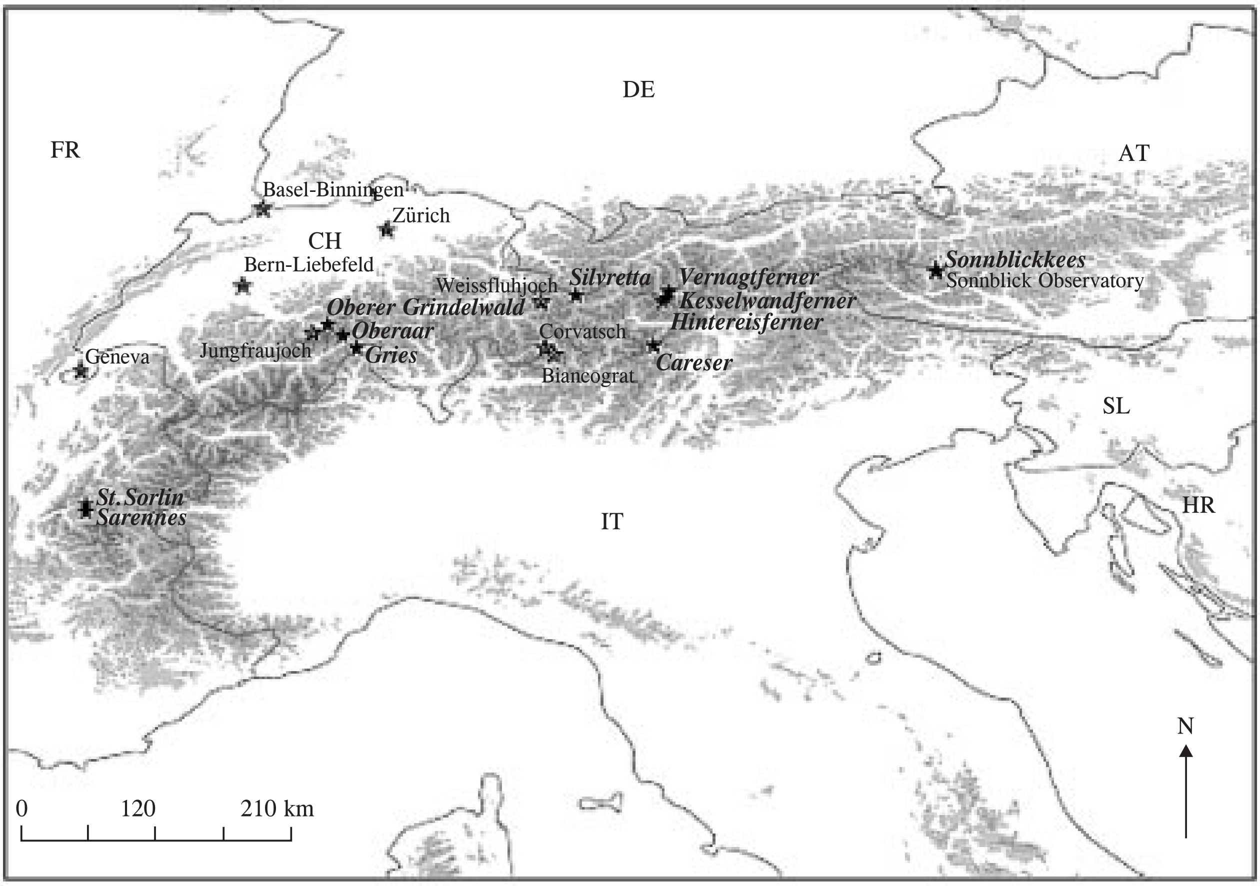

Fig. 5. Greater Alpine Region with the locations named in the text. Black stars represent glaciers. Countries labelled according to the ISO 2-digit country code. Country boundaries provided by ESRI. Background: European HYDRO1k-DEM, elevations above 1000 m are shaded in grey. Source: Land Processes Distributed Active Archive Center (LP DAAC), U.S. Geological Survey's EROS Data Center, http://edcdaac.usgs.gov

Рис. 5. Большой Альпийский регион, показаны места, упомянутые в тексте. Черными звездочками показаны ледники. Названия стран помечены в соответствии с двухзначным кодом стран ISO. Границы стран даны по данным ESRI. Фоновая карта: Европейская ЦМР HYDRO1k-DEM, области выше 1000 м оттенены серым цветом. Источник: Распределенный активный архивный центр по процессам суши (LP DAAC), Центр данных EROS Геологической службы США, http://edcdaac.usgs.gov

In the Alps, winter 2002/2003 started unusually early with snow fall down to 600 m by the end of September. In October the snow line shifted up- and downslope due to changing supply of warm and cold air mass into the Alps. November was characterized by a series of orographic upslope precipitation situations, bringing snow fall far above the average. At the Sonnblick Observatory (3107 m a.s.l., Austria) measured snow fall was three times higher than average, resulting in a total snow height of 2.8 m within 23 precipitation days [16].

After a fair weather period in January with temperatures of up to +4°C at 2000 m at the northern Alpine ridge, repeated snow falls resulted in large snow heights by the beginning of February. The amount of fresh snow fallen by February 6th exceeded the corresponding amount of the extraordinary winter 1999 (known in the Alps as “avalanche winter” due to the large amount of serious avalanche accidents). January and February were cold and predominantly poor in precipitation, especially south of the Central Main Alpine Ridge [16].

After February 6^{th}, no more snow fall events worth mentioning occurred until the beginning of April. February has been registered as the second or third most sunny February during the 103 years of measurements. According to snow records in Switzerland, March has been the month with the least amount of fresh snow since 1950. Sunshine duration in March reached 160–200% of the long-time mean. By mid-March a temperature rise of 12–15°C resulted in an uplift of the 0°C-isotherm to altitudes between 3100 and 3400 m. April started with the passage of a cold front bringing a cooling and snow fall down to the lowlands. With the intensive sunshine und the mild temperatures by the end of April snow melting and rising of the snowline continued rapidly. Normally, altitudes around 1500 m become snow free between mid- and end of May. In 2003 this happened already between mid- and end of April (Fig. 6 for evolution of the thickness of the snow pack in the Swiss Alps during spring 2003).

In the last days of April and the first ones of May, a second Sahara dust event occurred (the first one dating from mid-November). This dust, mixed with pollen and other pollution particles, led to an impressive discolouration of the snow.

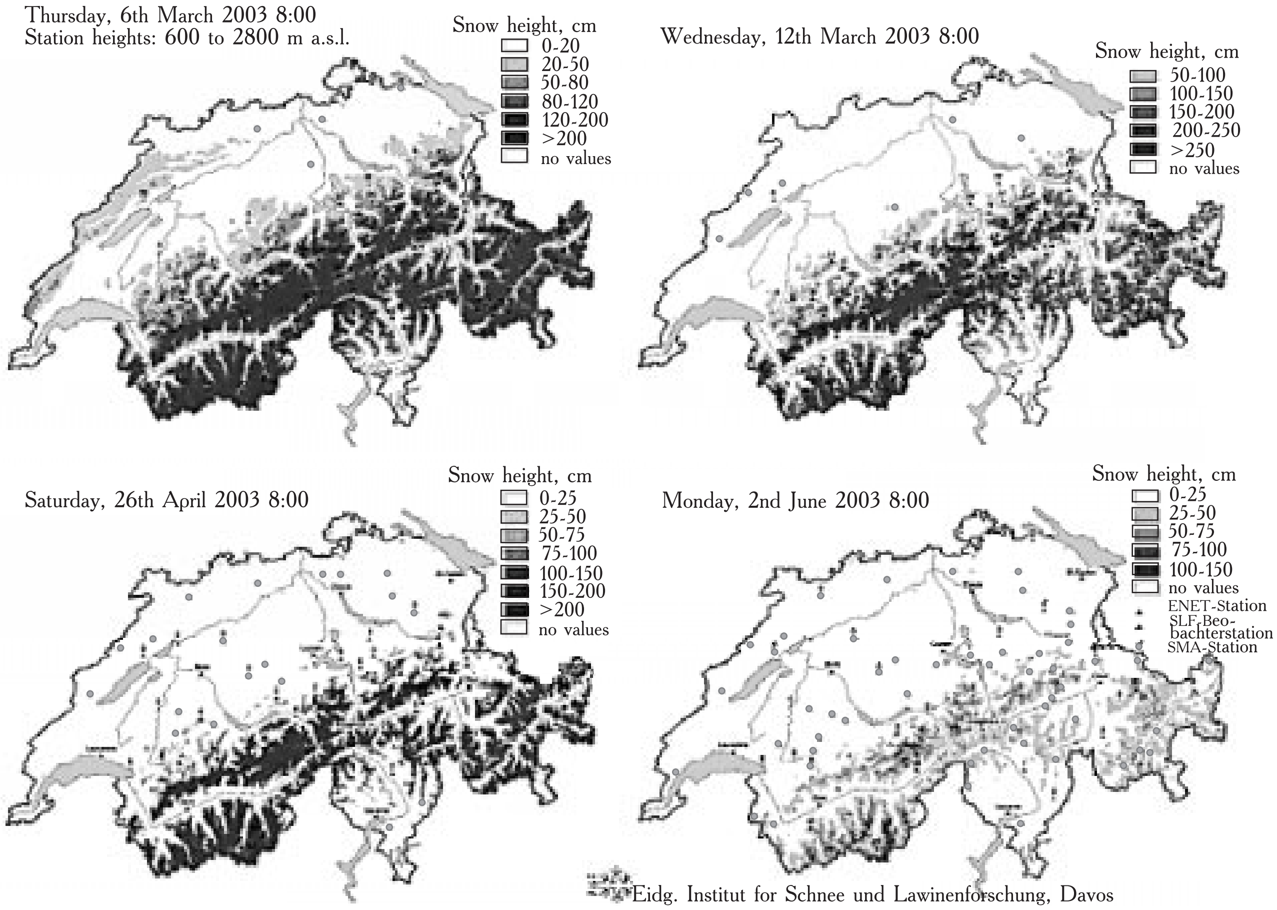

Fig. 6. Snow heights in Switzerland in March, April and June 2003 [19]. Snow heights are interpolated on actual terrain from automatic measurements and measurements by observing persons. Figures kindly provided by the Swiss Federal Institute for Snow and Avalanche Research

Рис. 6. Толщина снега в Швейцарии в марте, апреле и июне 2003 г.[19]. Толщина снега интерполирована по данным автоматических и ручных измерений с учетом реального рельефа местности. Рисунки любезно предоставлены Швейцарским федеральным институтом исследований снега и лавин

At the station Weissfluhjoch (2540 m a.s.l.), Switzerland, daily snow heights have been measured since 1936 (Fig. 7). In November 2002, total snow height reached almost the upper range of the corresponding long-term measurements. On February 8th 90% of maximum snow height of the past 67 years could still be found. The situation changed drastically during the following months: since 1936 there had been more snow at the end of May than in 2003 in 61 winters (rank 62 of 67). On the 1st of June only 34% of the corresponding long-term maximum snow depth were left. Finally, the observation site became snow free on the 14th of June. This is approximately six weeks earlier than on average.

After the unusually warm March and May, which in many regions had been the warmest May since the start of the measurements, the summer 2003 was marked by a record-breaking heat-wave, with its centre over France and Switzerland, that affected the whole European continent. The summer was dominated by anticyclonic influences, while clouds- and rain-bringing west winds rarely reached the Alps.

Fig. 8a [18] shows the temperature anomaly during the summer months June, July and August (JJA) with respect to the 1961–1990 mean, based on ERA-40 reanalysis data and operational meteorological analysis data. Monthly and seasonal temperature data from four stations in Switzerland (Basel-Binningen, Geneva, Bern-Liebefeld and Zurich) representative for the north-western foothills of the Alps were analyzed for the period 1864–2003. In 2003, temperatures in June, August and during the three summer months (JJA) were far off the distribution of 1864–2002 (Fig. 8b, d, e) [18]. The previous record holder for JJA was, for instance, 1947 with a temperature anomaly of $T'=2.7$°C (with respect to the 1864–2000 mean) [18]. The corresponding value for 2003 is $T'=5.1$°C and this amounts to an offset of 5.4 standard deviations from the mean [18]. Corresponding values of individual months are listed in Fig. 8b-e.

Record June temperatures at Sonnblick Observatory (3107 m a.s.l., Austria) were between 5.8 and 6.7°C above the mean of 1961–1990 [16]. In August, all daily mean temperatures at that location were above the norm and the monthly mean reached extraordinary 4.8°C [16]. In the Swiss Alps, recorded mean daily temperatures at Jungfraujoch (3580 m a.s.l.) amounted to 3°C and to 5.5°C at Corvatsch (2690 m a.s.l.) [19]. Using these values, calculated mean altitude of the 0°C-isotherm in July and August was around 4000 m.

The hot summer period ceased by the end of August when south-western winds brought warm and very moist air masses towards the Alps. After almost three months of drought heavy precipitation set in, propagated to the northern Alpine rim and lowered the snow line down to approximately 2000 m. The final end of the summer came by the beginning of October with a cold front from the west turning into a north-wind situation with orographic upslope precipitation, resulting in heavy snow fall down to 1000 m.

Overall, the year 2003 was 1.6 to 2°C warmer then the mean of 1961–1990. In some regions it was the warmest year that has ever been measured since the beginning of the records in 1880. With the extreme sunny months March and June, in the mountains also February, the yearly sunshine reached 115 to 130% of the mean of 1961–1990. It was the most sunny year since 1949, in some regions even since the beginning of the series in 1880. In addition, 2003 was one of the ten driest years of the past 103 years.

Glaciers. As described above, the winter snow cover was below average snow height. Already in May the snow was exposed to strong melting. Saharan dust, pollen and other pollution particles accumulated during ongoing melting and led to an albedo-feedback that enhanced the already strong ablation. Melting periods lasting for weeks without interruption resulted in extraordinary ablation values in 2003. Due to the early start of the ablation season, melting was not only strong but also of long duration. For the Vernagtferner in Austria, for example, it

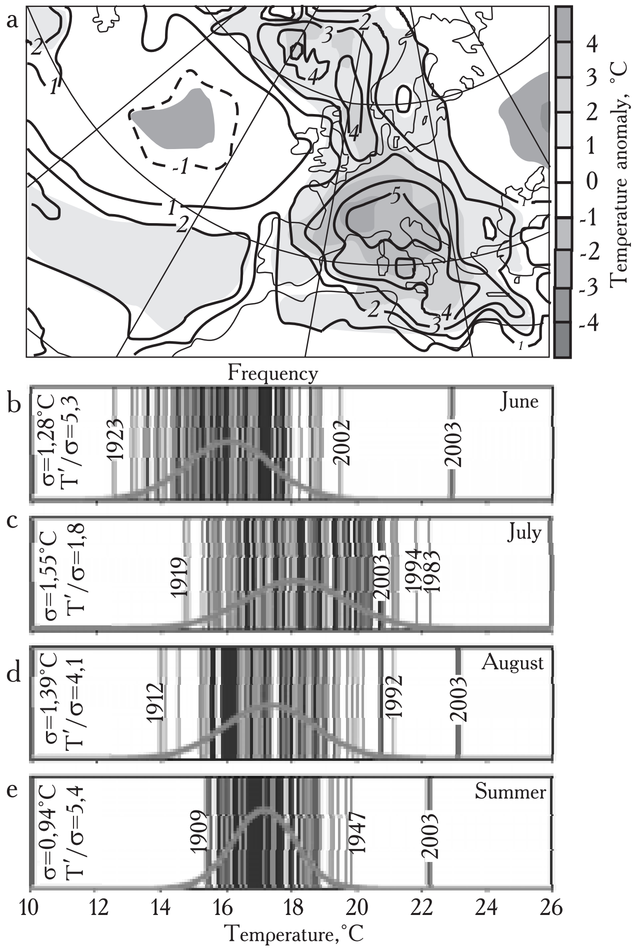

Fig. 8. Characteristics of the summer 2003 heat-wave [18]. a — JJA temperature anomaly with respect to the 1961–1990 mean. Shading shows temperature anomaly [°C], bold contours display anomalies normalized by the 30-year standard deviation. b–e) Distribution of Swiss monthly and seasonal summer temperatures for 1864–2003. The fitted gaussian distribution is indicated by the grey line. The values in the lower left corner of each panel list the standard deviation (σ) and the 2003 anomaly normalized by the 1864–2000 standard deviation (T'/σ). Figure reprinted with permission of the authors from [18]

Рис. 8. Характеристики волны летнего тепла 2003 г. [18]: a — температурная аномалия в июне-августе по отношению к среднему значению за 1961–1990 гг. Фоном разной интенсивности показана температурная аномалия, жирными изолиниями — аномалии, нормализованные по стандартному отклонению за 30 лет; b–e — распределения месячных и сезонных летних температур в Швейцарии за 1864–2003 гг. Подобранное гауссово распределение показано серой линией. Значения в нижнем левом углу каждого графика показывают стандартное отклонение (σ) и аномалию 2003 г., нормализованную по стандартному отклонению за 1864–2000 гг. (T'/σ). Рисунок публикуется с разрешения авторов [18]

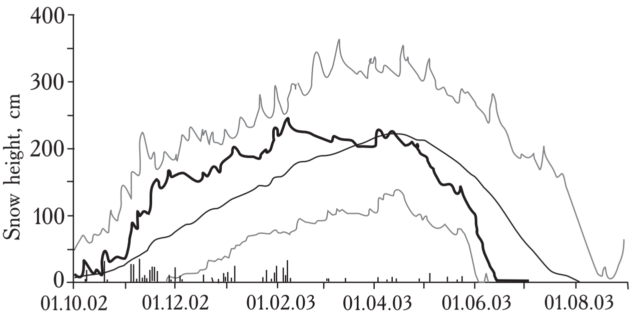

Fig. 7. Development of snow height during the hydrological year 2002/03 at the observation site on Weissfluhjoch (2540 m a.s.l.), Switzerland. Every morning snow height (bold line) and amount of fresh snow (black bars) are measured. Upper and lower lines mark the maximum and minimum snow heights since 1936, the grey line in the middle represents the long-term mean since 1936 [19]. Figure kindly provided by the Swiss Federal Institute for Snow and Avalanche Research

Рис. 7. Изменение толщины снега в течение 2002/03 г. на площадке наблюдений станции Вайсфлуйох (2540 м над ур. моря), Швейцария. Каждое утро измерялась толщина снега (жирная линия) и количество свежевыпавшего снега (черные столбики). Серые линии сверху и снизу показывают максимальную и минимальную толщину снега с 1936 г., а посредине — средние многолетние значения с 1936 г. Рисунок любезно предоставлен Швейцарским федеральным институтом исследований снега и лавин

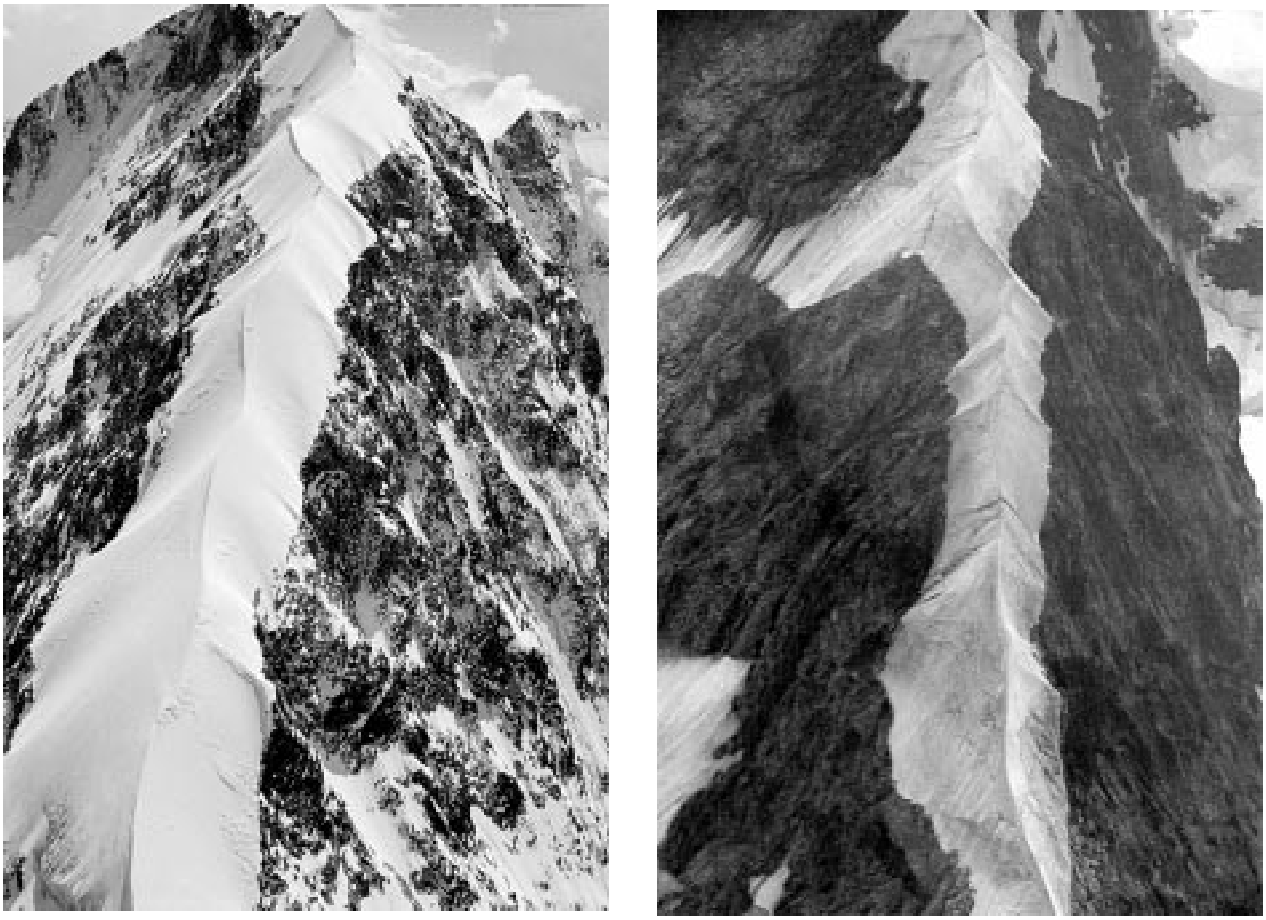

Fig. 9. Biancograt, Switzerland, reaching from approximately 3600 m (bottom of figures) to Piz Alv, 3995 m (top of figures). Left picture taken on August 17th 2002, right picture taken on August 9th 2003. Both pictures kindly provided by C. Rothenbühler

Рис. 9. Вид Бианкоград (Швейцария) в диапазоне высот примерно от 3600 м (нижняя часть фотографий) до 3995 м, Пиц-Алв (верхняя часть фотографий). Левый снимок сделан 17 августа 2002 г., правый — 9 августа 2003 г. Обе фотографии любезно предоставлены К. Ротенбюлером

lasted over 100 days, compared to the usual 50 to 60 days [20]. Ablation stake measurements at the tongues of Oberer Grindelwaldgletscher (1470 m a.s.l.) and Oberaargletscher (2340 m a.s.l.), Switzerland, showed an ice loss of approximately 13 m w.e. and 7 m w.e., respectively. Corresponding values in summer 2002 were 6–7 m w.e. at Oberer Grindelwaldgletscher and approximately 3 m w.e. at Oberaargletscher [11].

The snow line rose above the maximum elevations of many glaciers, leaving them without or only with minor accumulation areas. Fig. 9 shows the Biancograt in Switzerland, a ridge famous among mountain climbers. In August 2003 only its main ice ridge was left (Fig. 9, right). Snow, firn and ice that usually cover large parts of the neighbouring rock walls (Fig. 9, left) had largely melted away.

Reporting period for the «Glacier Mass Balance Bulletin, № 8 (2002–2003)» issued by the WGMS ends in fall 2004. Therefore, only preliminary mass balance data is available for the years 2002 and 2003. However, first rough estimates indicate an average loss in thickness of Alpine glaciers in 2003 of almost -3 m w.e. (Fig. 10).

Assuming a total Alpine glacierized area of 2909 km^2 in 1970/80 [5], relative loss in glacier size from 1973–1998/99 of 20% [17], an ice loss of almost -3 m w.e. in 2003 and a total ice volume of approx. 75 km^3 before 2003 (Paul, personal communication), estimated total glacier volume loss in 2003 corresponds to roughly 5–10% of the ice volume still present before 2003.

Discussion

Mass balances 1980–2001. As shown above, mean annual ice thickness loss of mountain glaciers is close to one third of a meter per year during the period 1980–2001. This results in a total thickness reduction of approximately 6 to 7 metres of ice since 1980. The spatially and regionally fairly well distributed sample of glaciers considered suggests that these values are, indeed, of worldwide representativity for mountain regions.

The spectacular loss in length, area and volume of mountain glaciers during the 20th century is a major reflection of the fact that rapid secular change in the energy balance of the earth's surface is taking place on a global scale [6]. The rate of this change is broadly consis-

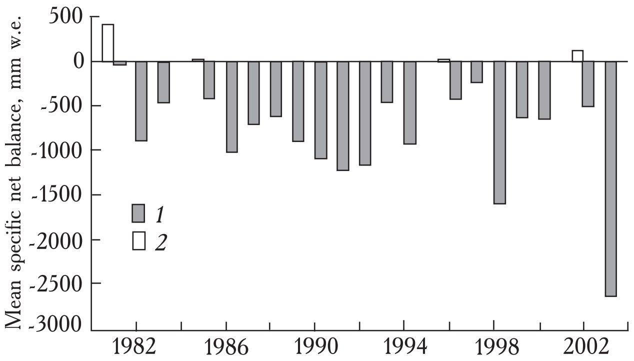

Fig. 10. Mean specific net balance of nine mountain glaciers in the European Alps for the period 1980—2003. The means are calculated using the annual mass balance of the following glaciers: St. Sorlin and Sarennes (France), Silvretta and Cries (Switzerland), Sonnblick, Vernagtferner, Kesselwandferner and Hintereisferner (Austria), Careser (Italy); 2002, 2003, incomplete sample, series subject to change; 1 — negative balances, 2 — positive balances

Рис. 10. Средний удельный чистый баланс 9 горных ледников Европейских Альп за 1980—2003 гг. Средние значения рассчитаны с использованием годовых величин баланса массы следующих ледников Альп: Сен-Сорлен и Саренс (Франция), Сильвретта и Грис (Швейцария), Сонбликс, Фернагтфернер, Кесельвандфернер и Хинтерайсфернер (Австрия), Кaресе (Италия). За 2002 и 2003 гг. данные неполны (ряды могут быть уточнены); 1 — отрицательные значения, 2 — положительные значения

tent with the estimated radiative forcing and changes in sensible heat as calculated with numerical climate models [6]. While the beginning of this rapid secular glacier retreat was probably little affected by human activity, there is little doubt that the current evolution contains an increasing anthropogenic influence [6].

Central Europe 2003. In the Alps, the beginning of 2003 was characterized by high accumulation lasting until the end of February. After that Alpine weather was dominated by anticyclonic weather types which are characteristic for high temperatures, high short-wave radiation, few clouds and low precipitation. This reversed the above-average accumulation from the first half of the winter into low snow heights in late winter, fast melting in spring and an unusually early start of the ablation season. The ablation period lasted for more than three months, again dominated by stationary anticyclonic systems over Europe. The snow line rose above the top of many glaciers, resulting in extensive melting of the firn layers. Together with the accumulated pollution particles this did not only enhance melting in 2003, but might also influence ablation in the coming years due to a reduced albedo of the glacier surfaces.

Mass balances of Hintereisferner and Sonnblickkees (both in Austria) and weather type classifications were analysed for the period 1966—1995 [8]. The author shows that for extremely negative years anticyclonic weather types dominated the accumulation and especial-

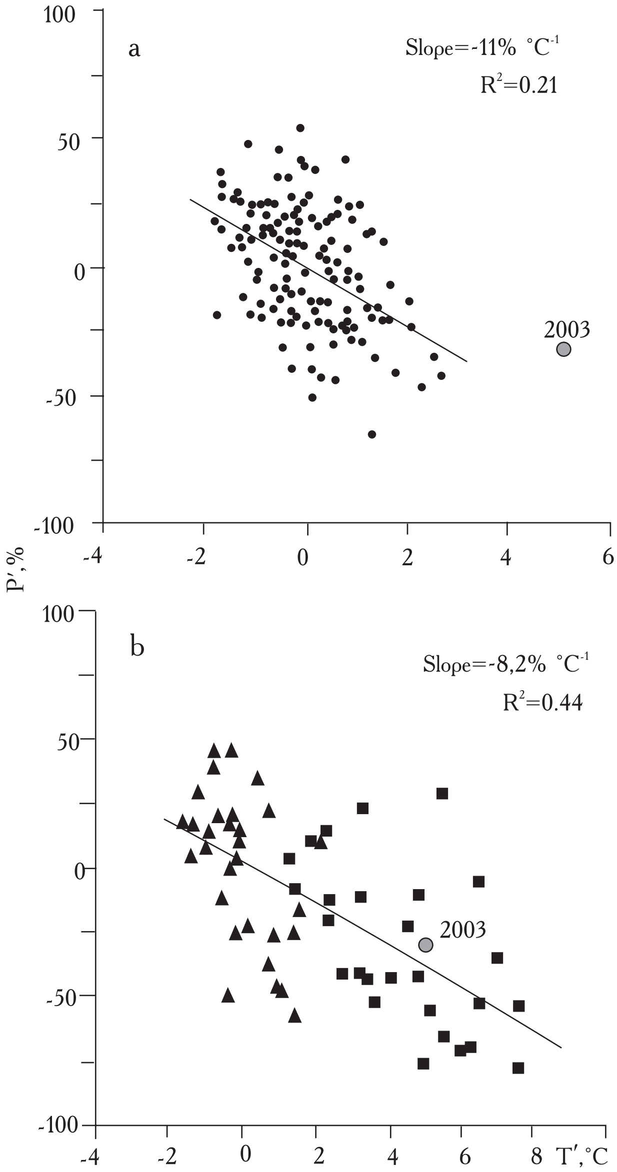

Fig. 11. Scatter diagrams showing summer temperature and precipitation anomalies for northern Switzerland. a) Long-term (1864—2003) station data with respect to 1961—1990 mean. b) Climate change simulation, control run (1961—1990, triangles) and scenario run (2071—2100, squares) with respect to control run mean. The bold circles show the observations for JJA 2003. The regression lines in a) and b) are based on 1864—2002 data and combined control and scenario run data, respectively. Figure reprinted with permission of the authors from [18]

Рис. 11. Диаграммы рассеяния, показывающие аномалии летних температур и осадков для северной Швейцарии: a — многолетние (1864—2003 гг.) данные по станциям, аномалии взяты по отношению к средним значениям за 1961—1990 гг; b — модельные данные климатических изменений, контрольный расчет (1961—1990 гг., треугольники) и сценарный расчет (2071—2100 гг., квадратики), приведенные к среднему значению контрольного расчета. Жирные кружки соответствуют наблюдениям за июнь-август 2003 г. Линии регрессии основаны на данных 1864—2002 гг. и объединенных данных контрольного и сценарного расчетов, соответственно. Рисунок публикуется с разрешения авторов [18] ly the ablation period of the two glaciers. Summer air temperature, winter precipitation and summer radiation balance may be used to parameterize glacier mass balance [e.g. 10, 15]. Such studies can help to understand inter-annual and regional dependence of glacier mass balance and general weather situations. However, to analyse intra-annual sensitivity of glacier mass balance to individual energy balance parameters the use of mass balance models is needed [e.g. 2, 9, 14].

The mean Alpine net balance in 2003 of almost -3 m w.e. (cf. above) is nearly twice as much as during the previous record year of 1998 (-1.6 m w.e.) and roughly five times more than the Alpine average loss of -0.6 m w.e. per year since 1980. In addition, this value is one order of magnitude higher than the global mean annual net balance of -0.3 m w.e. recorded during the already exceptionally warm period 1980–2001 (Fig. 3b, Figs. 4, 10).

The importance of the 2003 event in view of predicted climate change scenarios. The question if the extraordinary summer 2003 was only a maverick of today's climate or already a foreboding of things to come, is hard to answer. Therefore, possible future European climate was simulated by scientists from the ETH Zürich, using a regional climate model in a scenario with increased atmospheric greenhouse-gas concentrations and compared to results of long-term (1864–2003) station data from northern Switzerland, with respect to 1961–1990 (Fig. 11) [18].

Fig. 11 demonstrates that in terms of temperature and precipitation the climatic conditions in JJA 2003 were not unlike those simulated by the scenario run for the period 2071–2100. For northern Switzerland, the 2003 observation is located approximately in the middle of the scenario data points (Fig. 11b). Thus, the RCM simulations suggest that (under the given scenario assumptions) about every second summer could be as warm or even warmer (and dry or even dryer) than 2003 towards the end of the century [18]. Dramatic consequences of this scenario for glacier mass changes would have to be expected for the future. Volume losses in the order of 5–10% during one year (cf. above) would not constitute exceptional values but could occur frequently. Such developments would even exceed scenarios which are based on glacier shrinking rates of the last decades (25% volume loss in the last 25 years; loss of roughly two thirds of the original volume since 1850 [5]) that imply that less than 50% of the glacier volume still present in 1970/80 would remain in 2025 and only about 5–10% in 2100 [5].

Conclusion

Glacier mass balance, as the direct, undelayed signal of climate change, is among the most important parameters within global climate-related observation systems. Thus, efforts must be undertaken to resume interrupted series and to maintain and extend the currently existing worldwide observation network.

First results of the comparison of mass balance measurements from the extraordinary year 2003 and the Alpine and global mean implies that glacier melt continues at a considerable and most probably accelerating rate. The summer heat wave in 2003, mainly responsible for the extreme ice loss reported and analysed above, was an European event. However, this summer and its impact on glaciers may well serve as case study for a possible future climate and could therefore be of worldwide interest.

Post scriptum 27.1.2005

Meanwhile, the final values of the Alpine mass balances are available for all nine glaciers concerned [7]. The corrected values of the resulting mean specific net balance of the European Alps for 2002 and 2003 are -0.8 m w.e. and -2.5 m w.e., respectively. The value of 2003 is slightly smaller then the previously estimated value of nearly 3 m w.e.(see Text and Fig. 10). This does, however, not change the general conclusions that glacier melt in 2003 was extraordinary in the European Alps, and that the general decrease of global glacier mass continues at a fast, and probably accelerated rate.

Acknowledgements

We wish to thank C. Schär and his co-authors and the authors of the "Winter Reports" of the Swiss Federal Institute for Snow and Avalanche Research for providing us their figures. Thanks are also due to C. Rothenbühler for letting us use her outstanding photographs of the Biancograt. Further thanks go to ESRI for providing digital country contours and to our Russian colleagues for the translation of the abstract and the figure captions into Russian. We would also like to thank all national correspondents, principal investigators and contributing institutions of WGMS, without their work, our work would not be possible! The work of Michael Zemp is partly funded by the Swiss Federal Office of Education and Science (BBW-Contract 901.0498-2) within the EU programme ALP-IMP (Contract EVK2-CT-2002- 00148).

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Измерения баланса массы ледников мира: общие тренды и первые результаты исследований необычной ситуации 2003 г. в Центральной Европе

Ледники служат одним из ключевых индикаторов происходящих изменений климата. Помимо изменений длины ледников, где климатический сигнал отражается с запозданием, в фильтрованном и одновременно усиленном виде, особое значение имеют данные о балансе их массы, где этот сигнал отражается непосредственно и без какой либо задержки. Всемирная служба слежения за ледниками (Цюрих, Швейцария) собирает, стандартизирует и публикует каждые два года данные измерений баланса массы ледников мира. В настоящей статье рассматриваются тренды баланса массы ледников мира, в частности, за 1980–2001 гг. и для чрезвычайно жаркого и сухого лета 2003 г. в Центральной Европе, которое привело к потере 5–10% объема ледников Альп. Моделирование будущих температурных сценариев для северной Швейцарии дает основание полагать, что к концу текущего столетия каждое второе лето может быть столь же жарким или еще жарче (и таким же сухим или суше) [18]. Необычное лето 2003 г. наблюдалось в Центральной Европе, но такая ситуация может служить типичным примером возможного климата в будущем и в глобальном масштабе.

Paper IV

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Distributed modelling of the regional climatic equilibrium line altitude of glaciers in the European Alps

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Abstract

Glaciers are among the key indicators of ongoing climate change. The equilibrium line altitude is a theoretical line which defines the altitude at which annual accumulation equals the ablation. It represents the lowest boundary of the climatic glacierisation and, therefore, is an excellent proxy for climate variability. In this study we introduce a simple approach for modelling the glacier distribution at high spatial resolution over entire mountain ridges using a minimum of input data. An empirical relationship between precipitation and temperature at the steady-state equilibrium line altitude (ELA_0), is derived from direct glaciological mass balance measurements. Using geographical information systems (GIS) and a digital elevation model, this relationship is then applied over a spatial domain, to a so-called distributed modelling of the regional climatic ELA_0 ($rcELA_0$) and the climatic accumulation area (cAA) of 1971–1990 over the entire European Alps. A sensitivity study shows that a change in $rcELA_0$ of ±100 m is caused by a temperature change of ±1 °C or a precipitation decrease of 20% and increase of 27%, respectively. The modelled cAA of 1971–1990 agrees well with glacier outlines from the 1973 Swiss Glacier Inventory. Assuming a warming of 0.6 °C between 1850 and 1971–1990 leads to a mean $rcELA_0$ rise of 75 m and a corresponding cAA reduction of 26%. A further rise in temperature of 3 °C accompanied by an increase in precipitation of 10% leads to a further mean rise of the $rcELA_0$ of about 340 m and reduces the cAA of 1971–1990 by 74%.

Keywords: Glacier, Climate at Equilibrium Line Altitude, Climate Change, Geographical Information Systems

1. Introduction

Fluctuations of mountain glaciers are among the best natural indicators of climate change (IPCC, 2001). Glacier mass balance is the direct and undelayed signal of annual atmospheric conditions, i.e., temperature, precipitation and radiation balance (Haeberli, 2004). The equilibrium line altitude (ELA) on a glacier is a theoretical line which defines the altitude at which annual accumulation equals the ablation. It represents the lowest boundary of the climatic glacierisation – that is, where the glacierisation can begin. Thus, climatic variability can be inferred from variations in the ELA (Kuhn, 1981; Ohmura et al., 1992; Lie et al., 2003a).

To understand the climate at the ELA the processes of accumulation and ablation can be examined, the latter based on the energy balance principle. Another approach is to search for relationships between the relevant parameters for the ELA and the glacier accumulation or ablation (Ahlmann, 1924; Shumsky, 1964; Ohmura et al., 1992).

Physical models describing the relationship between glacier behaviour and climate change can be very sophisticated (Greuell, 1989; Klok and Oerlemans 2002), in particular those which include the mechanics of glacier flow, driven by the mass balance (e.g., Leysinger-Vieli and Gudmundsson, 2004). These models require detailed input data from various sources. Because of their complexity, they facilitate a deeper understanding of physical glacier behaviour (Kerschner, 2002). However, due to the large amount of required input data, they can be applied only to recent glacier states and to single glaciers or glacierised catchments. Other studies use correlations between glacier energy balance, ablation and meteorological parameters to complete and extend mass balance series in time (e.g., Hoinkes and Steinacker, 1975; Braithwaite, 1981; Letréguilly and Reynaud, 1990).

Our aim is a) to present a simple approach that is able to model the glacier distribution at high spatial resolution (about 100 m) over entire mountain ridges, b) to investigate the sensitivity of the Alpine[1] glacierisation to changes in temperature and precipitation, and c) to demonstrate the possible impact of a climate scenario on Alpine glacierisation in the decades to come. To achieve this, an empirical relationship between precipitation and temperature at the steady-state equilibrium line altitude (ELA_0) is derived from direct glaciological mass balance measurements. GIS techniques and a DEM are then used to apply this relationship over a spatial domain to a so-called distributed modelling of the regional climatic ELA_0 ($rcELA_0$), and the climatic accumulation area (cAA) of 1971–1990 over the entire European Alps. Different sources of model errors and the order of magnitude of their amounts are discussed. In conclusion, a sensitivity study demonstrates the potential of this distributed modelling approach for the reconstruction of historical $rcELA_0$ and future scenarios.

2. Previous studies

In the relevant literature, and in the various languages used there, no consistent terminology can be found for the equilibrium line altitude. Gross et al. (1978) gave an overview of definitions of glacier snowline, firn line, equilibrium line and different approaches for estimating these. In this study we use the original terms cited in the literature. In addition, we distinguish between equilibrium line altitude (ELA), steady-state equilibrium line altitude (ELA_0), regional climatic equilibrium line altitude ($rcELA_0$) and local topographic equilibrium line altitude ($ltELA_0$).

[1] In this article ‘Alps’ or ‘Alpine’ refer explicitly to the European Alps, the terms ‘alps’ or ‘alpine’ are purely generic.

Haeberli (1983) presented a schematic diagram of glacier limits after Shumsky (1964), describing the distribution of glaciers primarily as a function of mean annual air temperature and annual precipitation. As a general rule, in humid-maritime regions, the ELA is at low altitude because of the great amount of ablation required to eliminate the deep snowfall. Temperate glaciers dominate these landscapes. Such ice bodies, with relatively rapid flow, exhibit a high mass turnover and react strongly to atmospheric warming by enhanced melt and runoff. Under dry continental conditions the ELA is at high elevation. In such regions, glaciers are polythermal or cold and feature a low mass turnover.

The basic idea of discovering an empirical relationship between precipitation or temperature and the snowline goes back to Ahlmann (1924). Krenke (1975) pursued a similar concept for estimating precipitation in high mountain areas in the former Soviet Union. Ananicheva and Krenke (2005) used such relationships to investigate the evolution of the climatic snow line and the ELA in north-eastern Siberia Mountains. Gross et al. (1978) suggested a steady-state Accumulation Area Ratio (AAR_0) of 0.67 for glaciers in the Alps as an approximation of the ELA_0 and zero mass balance and compared it to the eight most frequently applied methods for the estimation of the snowline. Furthermore they used the AAR_0 to estimate the snowline depression of Late-Würm (20,000–10,000 years BP) glacier readvances. With the AAR_0 method, the ELA_0 of past glaciers is fairly easy to determine from topographical glacier maps and hence often applied in palaeo-climatic studies (e.g., Kerschner, 1985; Maisch, 1992; Kerschner et al., 2000; Maisch et al., 2000). However, AAR_0 values obtained from worldwide mass balance measurements vary between 0.22 and 0.72 (IUGG(CCS)/UNEP/UNESCO/WMO, 2005). Therefore, the AAR_0 method should not be used uncritically.

Multivariate regression models, relating the change in the dependent variable (ELA) to the changes in the independent variables (climate data) were presented by Ohmura et al. (1992), Kerschner (1996) and Greene et al. (1999). The results of the multivariate regression analysis are difficult to interpret from a physical point of view and, for statistical reasons, are of only limited value for extrapolation beyond the spatial and temporal limits of the independent variables in the original data set (Kerschner, 1996).

Shea et al. (2004) modelled a simple cell-based climatology of the Canadian Rockies from temperature, precipitation and snowfall lapse rates constructed from climate normal, and elevation attributes extracted from a 1-kilometer DEM for the period 1961–1990. Fractional glacierised areas were calculated from a digital glacier inventory for the 0.25° x 0.5° model cells. Single and multivariate statistical analysis was then performed to determine the climatic variables controlling the spatial distribution of glaciers in the Rockies.

Lie et al. (2003a) presented an exponential relationship between mean ablation-season temperature and winter precipitation at the ELA of ten Norwegian glaciers. Using interpolated precipitation and temperature data for the period of 1961–1990 and a DEM of southern Norway, they modelled the altitude of instantaneous glacierisation (AIG) and the corresponding glacier accumulation areas, with a spatial resolution of 5 degree minutes (Lie et al., 2003b).

The approach presented in this study is based on the concept of Haeberli (1983) and Shumsky (1964). We present an empirical relationship between 6-month summer temperature and annual precipitation at the ELA_0, based on the period of 1971–1990. Rather than applying the often-used AAR_0 method, we determine the ELA_0 from direct glaciological mass balance measurements. Implementing the approach from Lie et al. (2003a), we then apply the obtained empirical relationship, using GIS techniques, to model the AIG, the rcELA$_0$ and the cAA with a spatial resolution of 3 arc seconds (approx. 100 m) over the entire European Alps.

3. Methods

In Fig. 1 an overview is given of all the elements mentioned later in the article, such as locations of mass balance observations, high altitude weather stations, hydrological basins and extent of the Alpine mountain ridge.

3.1. Precipitation and temperature at the ELA$_0$

To set up an empirical relationship between precipitation and temperature at the ELA$_0$ we focus on the period of 1971–1990. In this period, mass balance series from 14 Alpine glaciers as well as Alpine-wide temperature and precipitation data are available. The ELA$_0$ is calculated for each glacier from the relation between the specific net balance and the ELA, as compiled by the World Glacier Monitoring Service (http://www.wgms.ch) and published in the Glacier Mass Balance Bulletin series (e.g. IUGG(CCS)/UNEP/UNESCO/WMO, 2005). ELA values outside the glacier altitude range (i.e., when the entire glacier is accumulation or ablation area) are excluded in this regression analysis. Fig. 2 shows the example of Vernagt Glacier (Austria) with an ELA$_0$ of 3,076 m a.s.l. The regression analysis of the 14 glaciers show coefficients of determination (R^2) ranging from 0.74 to 0.99, with a mean of 0.89.

Precipitation values at the glacier ELA$_0$ were obtained from the Alpine precipitation climatology (1971–1990) published by Frei and Schär (1998) and Schwarb et al. (2001). Based on a comprehensive database with observations from 5,831 conventional rain gauges and 259 totalisators (i.e. cumulative precipitation gauges), this gridded data set provides mean monthly precipitation (1971–1990), as well as monthly precipitation-elevation gradients on a spatial resolution of 1.25 arc minutes (approx. 2 km). Frei and Schär (1998) and Schwarb et al. (2001) calculated a regression between precipitation and altitude for each grid cell. Station data is weighted differently according to the extent to which they are representative of conditions at that grid cell. The weighting takes into account factors such as distance and difference in altitude from the grid cell, as well as varying aspect. However, in this study we find that the DEM that underlies the precipitation climatology differs significantly in altitude from the ELA$_0$, i.e., the altitude assigned to a grid cell in the precipitation climatology is significantly different from the measured ELA$_0$ (mean absolute difference of 161 m, with a maximum absolute difference of 655 m). Therefore, we use the precipitation-elevation gradients (ranging at ELA$_0$ locations from -2 to 78% per 100 m) to correct the precipitation of each grid cell in the glacier accumulation area to the corresponding ELA$_0$. Not knowing the exact location of the ELA$_0$ (only the location of the glacier boundaries and the altitude value of the ELA$_0$ are known) and of the uncertainties of the precipitation climatology, the precipitation at the ELA$_0$ is calculated as the mean of all precipitation grid cells covering the glacier accumulation area.

The variations in precipitation are complex and, in contrast to temperature variations, can be poorly correlated with altitude. Auer et al. (2005) show that the spatial mean decorrelation distances (R^2 decreasing below 0.5) of temperature (annual: 993 km, seasonal: 765 km, monthly: 722 km) are much larger than those of precipitation (annual: 149 km, seasonal: 120, monthly: 105 km). Hence, many fewer stations are needed to interpolate temperature at glacier locations. However, Oerlemans (2001, p. 37–38) demonstrates for the Morteratsch Glacier (Switzerland) that temperature at a nearby valley station may tend to become decoupled from the temperature measured at a automatic weather station located on the glacier surface (e.g., due to atmospheric inversions in the flat valley during winter), whereas the correlation with a high altitude weather station remains high. Therefore, we use temperature from 12 Alpine high altitude weather stations with continuous data series between 1971 and 1990, provided by MeteoSwiss, Zurich, and the Central Institute for Meteorology and Geodynamics, Vienna. The 12 stations are all located above 2000 m a.s.l., with the exception of Feuerkogel (Austria) at 1,618 m a.s.l. Monthly lapse rates are empirically derived from temperature and elevation values of the 12 stations. All of these linear regression analysis show high correlations (R^2 above 0.9) and range from 0.54 °C per 100 m in December to 0.71 °C per 100 m in July. Seasonal mean temperatures and lapse rates are calculated. With these empirically derived lapse rates, seasonal temperatures are altitude-adjusted to 2,000 m a.s.l. Using the inverse distance weighted (IDW; Watson and Philip, 1985) interpolation method, temperatures are interpolated over the entire Alps. Anisotropy is taken into account by specifying a lower power along and a higher power against the Alpine ridge (specifying a lower value for power will provide more influence to surrounding points farther away). Fig. 3 shows the altitude-adjusted 6-month summer temperature field at 2,000 m a.s.l. over the greater Alpine region. Temperature ranges from 5.0–7.1 °C. At glacier locations the empirical lapse rates are used again to extrapolate temperatures at 2,000 m a.s.l. to the ELA_0. The same principle is applied to produce an Alpine-wide data set of temperature on terrain, with a resolution of 3 arc seconds (approx. 100 m), by extrapolating the temperature field at 2,000 m a.s.l. with the seasonal lapse rate to the altitude of a DEM (Fig. 4). As already mentioned, the exact location of the ELA_0 is not known and it might also differ from the elevation of the corresponding grid cell of the DEM. For the estimation of temperature and precipitation at the ELA_0, used to derive the empirical relationship, we consider, therefore, the ELA_0 instead of the corresponding altitude of the DEM. Hence, temperature and precipitation values at the location of the mass balance observations might differ from the values at the corresponding locations in the gridded precipitation and temperature climatologies.

Annual and 6-month winter precipitation, and annual and 6- and 3-month summer temperatures at the ELA_0 of the 14 Alpine glaciers are shown in Table 1. The empirical relationship between annual precipitation (P_a) and the 6-month summer temperature (T_{A-S}) at the ELA_0 performs best with an R^2 of 0.49 on a 1% significance level (p-value of 0.0052) and can be expressed by the regression equation:

$$P_a = 1773 \cdot e^{0.2429 \cdot T_{A-S}} \tag{1}$$

Equation (1) is plotted in Fig. 5 together with the upper and lower boundaries of the 95% confidence interval which can be formulated as:

$$P_a = 1773 \cdot e^{(0.2429 \cdot T_{A-S} \pm 0.1985 \cdot \sqrt{(1+(T_{A-S}+0.2301)^2})} \tag{2}$$

These boundaries describe the possible range of annual precipitation, at a 5% significance level, for a given 6-month summer temperature at the ELA_0.

The 95% confidence interval boundaries can be approximated by a shift of Equation (1) in such a manner that the new equations for the lower boundary, Equation (3), and upper boundary, Equation (4), describe a confidence interval greater than 95% within the temperature data range and lower than 95% outside the temperature data range, respectively.

$$P_a = 0.7 \cdot 1773 \cdot e^{0.2429 \cdot T_{A-S}} \tag{3}$$

$$P_a = 1.45 \cdot 1773 \cdot e^{0.2429 \cdot T_{A-S}} \tag{4}$$

3.2. Distributed modelling of the regional climatic ELA$_0$

To apply Equation (1) in space, we implement the approach from Lie et al. (2003a). Along with this, we introduce temperature lapse rate and precipitation-elevation gradient to derive temperature and precipitation at the AIG. Temperature at the AIG (T_{AIG}) can be derived from the temperature at a known altitude (T_0), the lapse rate (ΔT) in degrees Celsius per 100 m and the height of the AIG above the terrain (h, in 100 m units):

$$T_{AIG} = T_0 - (\Delta T \cdot h) \tag{5}$$

The precipitation at the AIG (P_{AIG}) can be derived from the precipitation at a known altitude (P_0), the precipitation-elevation gradient (ΔP) in percent per 100 m and the height of the AIG above the terrain (h, in 100 m units):

$$P_{AIG} = P_0 \cdot (1 + \Delta P)^h \tag{6}$$

By combining Equations (5) and (6) with Equation (1), the expression becomes:

$$P_0 \cdot (1 + \Delta P)^h = 1773 \cdot e^{0.2429 \cdot (T_0 - (\Delta T \cdot h))} \tag{7}$$

Solving Equation (7) with respect to the only unknown parameter h, the AIG is:

$$AIG = ALT_{DEM} + (h \cdot 100) = ALT_{DEM} + \left(\frac{\ln(1773) + 0.2429 \cdot T_0 - \ln(P_0)}{\ln(1 + \Delta P) + 0.2429 \cdot \Delta T} \cdot 100\right) \tag{8}$$

where ALT_{DEM} is the altitude of the terrain. The rcELA$_0$ can now be computed from the intersection of the DEM with the AIG, where ALT_{DEM} is equal to AIG (or: h = 0). The climatic accumulation area (cAA) corresponds to the regions where the ALT_{DEM} is above the AIG (or: h < 0):

$$AIG \begin{cases} < & ALT_{DEM} & \rightarrow & cAA \\ = & ALT_{DEM} & \rightarrow & rcELA_0 \\ > & ALT_{DEM} & \rightarrow & no.cAA \end{cases} \tag{9}$$

A schematic representation of the AIG and its association with ltELA$_0$, rcELA$_0$ and cAA appears in Fig. 6.

The AIG, rcELA$_0$ and cAA are modelled in space using the annual precipitation data (in mm) and the corresponding annual precipitation-elevation gradients (in percent per 100 m) from Frei and Schär (1998) and Schwarb et al. (2001) and 6-month summer lapse rate (0.67 °C per 100 m) and 6-month summer temperature on terrain as input raster data sets. The summer temperature data set is extrapolated with the summer lapse rate from the altitude-adjusted temperature field at 2,000 m a.s.l. (Fig. 3) to the altitude of the DEM (Fig. 4). The latter is the 3 arc second version created by the Shuttle Radar Topography Mission (SRTM3; spatial coverage from 61° N to 57° S) and resampled to 100 m spatial resolution for our purposes. The data can be downloaded for free from a NASA ftp-server (ftp://e0mss21u.ecs.nasa.gov/srtm/) and are described in Rabus et al. (2003). Larger gaps in the SRTM3 data set have been filled by resampled 30 arc second SRTM30 elevation data, and can be obtained from the same source. The distributed modelling is implemented and automated using the ESRI software ArcGIS 9.0. A model run takes a few minutes of computing time.

In Fig. 7 the modelled cAA is shown for the reference period 1971–1990, and the maximum spatial extent of the investigation area (6–14° E, 44–48° N) is defined by the data set with the minimum spatial extent (precipitation data). By combining Equations (5) and (6) with Equations (3) and (4), AIG, $rcELA_0$ and cAA of the boundaries of the approximated confidence interval can be applied in space as well (Fig. 7, inset map).

3.3. Sensitivity study

In addition to the model run for the reference period 1971–1990, six more model runs were carried out to study the sensitivity of the ELA_0 to changes in temperature and precipitation. Temperature and/or precipitation is altered by a uniform deviation over the entire investigation area (i.e., the entire gridded climatologies). Table 2 gives an overview on the seven model runs.

MR_{T+1} and MR_{T-1} as well as MR_{P+25} and MR_{P-25} are sensitivity studies with deviations only in temperature or precipitation. $MR_{T-0.6}$ represents a summer temperature cooling of 0.6 °C, as assumed by Maisch et al. (2000) for the year 1850. $MR_{T+3/P+10}$ applies a warming of the summer temperature of 3 °C and a concurrent rise in precipitation by 10%. This corresponds to a moderate climate change scenario as published by OcCC (2004), commissioned by the Swiss Federal Office of Energy (SFOE). This study is based on IPCC (2001) and estimates a summer (JJA) temperature rise of 0.8–5.1 °C by 2050 for the northern and 1.0–5.6 °C for the southern slope of the Swiss Alps, respectively. It estimates precipitation to decrease in summer and increase in winter. For winter (DJF) precipitation a rise of 5–23% is indicated by 2050.

Reference and sensitivity model runs are analysed for individual glacier regions, within the hydrological basins, as derived from the HYDRO1k DEM (provided by the Land Processes Distributed Active Archive Centre located at the U.S. Geological Survey's EROS Data Centre: http://LPDAAC.usgs.gov), and within the 1973 outlines of the Swiss Glacier Inventory (Kääb et al., 2002; Paul et al., 2002) for glaciers larger than 1 km^2.

4. Results

4.1. Empirical relationship between precipitation and temperature at the ELA_0

Precipitation at the glacier ELA_0 (Equation 1, Table 1) shows a non-linear increase with increasing temperature. The differences in R^2 between the applied exponential trends and linear trends range from 0.06–0.08, i.e., 6–8% of the explained variance. The empirical relationships with annual precipitation perform better than the ones with 6-month winter precipitation, with R^2 differences of 0.10–0.12. For the distributed modelling of the AIG, the $rcELA_0$ and the cAA the relationship between annual precipitation and 6-month summer temperature is used (Equation (1)), as it performs best with an R^2 of 0.49 on a 1% significance level (p-value of 0.0052).

4.2. Modelled glacierisation of the reference run (1971–1990)

The AIG of the reference model run (1971–1990), shows two culminations with AIG values above 3,100 m a.s.l.; one ranging from the southern Valaisan Alps, Switzerland, to the south-western Alps and another one between the Rhaetian Alps, Switzerland, over the South Tyrol, Italy, to the Upper Tauern, Austria. There is an AIG depression over the Ticino Alps, Switzerland, and a generally lower AIG on the northern Alpine slope.

Table 3 lists the $rcELA_0$ averaged within the hydrological basins and within the Swiss glacier outlines. As the $rcELA_0$ represents the intersection of the AIG with the DEM, it reproduces the general image of the AIG. It ranges from 2,591 m a.s.l. (Isar) to 3,183 m a.s.l. (Durance) with a mean value of 2,951 m a.s.l. However, some hydrological basins show an average, caused by local topographic effects, that does not correspond to the $rcELA_0$ of the sub-regions. The most pronounced example is the Rhone basin in the Valais, Switzerland. It has an average ELA_0 of 2,966 m a.s.l. with a standard deviation of 161 m. This is produced by $rcELA_0$ values below 2,800 m a.s.l. in the northern Valais and above 3,100 m a.s.l. in the southern Valais.

Over the entire Alps the cAA covers an area of 3,059 km^2. As the cAA is simply the terrain above the $rcELA_0$, it does not distinguish between glacier surface and ice-free rock walls. In a first order approach this can be taken into account by applying a slope-dependent glacier fraction to the modelled cAA. Table 4 shows the slope-area distribution (derived from the SRTM3) within the cAA and within the parts of the 1973 outlines of the Swiss Glacier Inventory inside the cAA of a test area (Canton Valais, Switzerland, corresponding approximately to the Rhone basin). From these parameters, glacier fractions are calculated for the nine slope classes and applied to the slope-area distribution of the Alpine cAA. These empirical glacier fractions are dependent on quality and resolution of the DEM used and therefore have to be re-evaluated when using other DEMs. The corrected cAA equals 1,950 km^2 and corresponds to an AAR_0 of 0.67 of the measured total Alpine glacier area in the 1970s, which was 2,909 km^2 (Zemp et al., in press).

4.3. Model validation

The model error can be described by the deviations of the modelled AIG values of the 14 glaciers from the measured ELA_0. The mean absolute difference is 66.7 m with a standard deviation of 55.4 m. The greatest deviations are found for Silvretta Glacier, where the measured ELA_0 is 214 m below the modelled AIG, and for Grosser Aletsch Glacier, where the ELA_0 is 127 m above the AIG. To get an estimation of the possible deviation of the $ltELA_0$ from the $rcELA_0$ (as modelled), Equation (1) is substituted by the two approximations of the confidences interval boundaries, Equations (3) and (4), to calculate a spatial confidence zone. This results in average deviations of -177 m and +209 m from the reference $rcELA_0$, based on Equation (1).

The resulting cAA is checked in a qualitative manner by comparison with the outlines from the 1973 Swiss Glacier Inventory and with LandsatTM images dating from the 1990s. The modelled cAA corresponds well overall to the real accumulation areas of Alpine glaciers and there are minor quantities of cAA cells in regions with no glacierisation. A general overestimation of the accumulation areas on SE–SW slopes and underestimation on NE–NW slopes can be found.

4.4. Sensitivity study

The $rcELA_0$ deviations of the sensitivity model runs from the reference run (1971–1990) within the hydrological basins are shown in Table 3. A temperature change of ±1 °C leads to an average $rcELA_0$ deviation of +137/-125 m, ranging from +112 m (Aosta) to +190 m (Isar) and from -44 m (Vorderrhein) to -201 m (Var), respectively. A precipitation change of ±25% leads to an average $rcELA_0$ deviation of -114/+157 m, with a range similar to the one-degree temperature deviation. $MR_{T-0.6}$ results in an average $rcELA_0$-decrease of 75 m, ranging from 24 m to 131 m. The increase of 18 m within the Mur basin is an undesired side-effect caused by the very small number of cAA cells. The occurrence of an additional glacierised peak in the $MR_{T-0.6}$ run leads to an average $rcELA_0$ higher than the one from the reference model run (1971–1990). The total cAA of the $MR_{T-0.6}$ run amounts to 4,157 km^2. $MR_{T+3/P+10}$ leads to an average $rcELA_0$ rise of 336 m and the disappearance of glaciers in eight out of 28 basins. The corresponding total cAA over the entire Alps shrinks to 812 km^2. When the slope-dependent glacier fraction (Table 4) is applied, the cAA of the $MR_{T-0.6}$ and the $MR_{T+3/P+10}$ amounts to 2,650 km^2 and 504 km^2, respectively. The average $rcELA_0$ and the deviations corresponding to the sensitivity runs, calculated within the 213 Swiss glacier outlines, represent a $rcELA_0$ subset of the hydrological basins Aare, Adda, Hinterrhein, Inn, Landquart, Linth, Reuss, Rhone, Toce and Vorderrhein. The deviations within the glacier outlines are slightly more moderate than those of the corresponding basin. The $MR_{T-0.6}$ induced a $rcELA_0$ lowering of 65 m.

Figures 8 and 9 show the modelled cAA of the reference run (1971–1990) and of $MR_{T+3/P+10}$ for the regions of Grosser Aletsch Glacier and Rhone Glacier. To facilitate a qualitative check, the outlines of the 1973 Swiss Glacier Inventory are shown as well. The modelled cAA of the reference run (1971–1990) covers the accumulation areas of the Swiss glacier outlines of 1973 reasonably well. In flat areas (e.g., upper tongue of Grosser Aletsch Glacier, Fig. 8) the coarser resolved precipitation data may determine the $rcELA_0$ (i.e., long, sharp boundaries), whereas on steeper terrain, the temperature data set with higher resolution (20 times) is the determining factor. In general, cAA is overestimated on SE–SW slopes and underestimated on NE–NW slopes. The $rcELA_0$ rise between the reference run and $MR_{T+3/P+10}$ leads to a much more pronounced cAA reduction in flat regions compared to steep slopes.

In regions where the glacier outlines are available, the rise of the $rcELA_0$ can be derived for individual glaciers. In Fig. 8, the rise of the $rcELA_0$ between the two model runs amounts to 249 m for Grosser Aletsch Glacier, 315 m for Fiescher Glacier, 314 m for Oberaar Glacier, 310 m for Gauli Glacier and 353 m for the Oberer Grindelwald Glacier. In Fig. 9 the $rcELA_0$ rises by 384 m at Rhone Glacier, 381 m at Damma Glacier and 374 m at Trift Glacier. In this region the temperature rise of the $MR_{T+3/P+10}$ leads to a strong reduction and disintegration of the cAA.

5. Discussion

5.1. Error estimations and model uncertainties

The 14 glaciers, on which the relationship between precipitation and temperature at the ELA_0 is based, are geographically and climatologically well-distributed over the Alps. Only the south-western Alps between France and Italy lack temperature and mass balance data. In this region, temperature measurements from high altitude weather stations began only after 1990. Mass balance data from Saint Sorlin (45° 11' N, 6° 10' E) and Sarennes (45° 7' N, 6° 10' E) exist, but ELA values are not available.

The precipitation data set by Frei and Schär (1998) and Schwarb et al. (2001) is based on more than 6,000 data points and is currently the best available data set at this resolution covering the entire Alps. The applied interpolation method takes into account the fact that the precipitation-elevation gradients can vary considerably from one region to another (Schwarb et al., 2001). Approaches using a constant precipitation-elevation gradient for the precipitation extrapolation to the ELA may work in regions dominated by orographic precipitation due to frontal cyclones flowing perpendicular to the orientation of the mountain ridge (e.g., western side of Canadian Rockies, western mountains in southern Norway), but will not include the effects of the complex gradient-patterns of the European Alps. However, in the precipitation climatology used, the data are not adjusted to the systematic precipitation measurement error, since the data needed to do this (wind, aspect of the station) are not available. Sevruk (1985) estimates this error for mean annual precipitation at 15–30% above 1,500 m a.s.l. Hence, Equation (1) has to be adapted if it is applied or compared to precipitation data with corrected systematic measurement error.

As already mentioned in the Methods section, the spatial decorrelation distance is much greater for temperature than for precipitation. Together with the high coefficients of determination ($R^2 > 0.9$) of the derived temperature lapse rates, this justifies the interpolation of the general high altitude temperature field (at 2,000 m a.s.l.) over the Alps based on the data from only 12 stations. Comparisons of the applied IDW interpolation method (considering anisotropy) with IDW (not considering anisotropy) and KRIGING (Oliver and Webster, 1990) show a mean temperature difference of 0.04 °C (with a standard deviation of 0.16 °C) and -0.03 °C (with a standard deviation of 0.35°), respectively. Assuming a temperature lapse rate of 0.67 °C per 100 m this corresponds to a mean ELA difference of about 5 m. In a test area of about 42 km x 34 km in the southern Valaisan Alps, Switzerland, comparisons of the SRTM3 (used to extrapolate temperature at 2000 m a.s.l. to the terrain) with the DEM25L2 (a high-quality Swiss DEM with a resolution of 25 m) show a mean altitude difference between the two DEMs of 7.7 m with a standard deviation of 53.7 m. In the regions where the SRTM3 gaps had to be filled with the SRTM30, the DEM difference amounts to 100 m with a standard deviation of 158 m. This shows that the temperature error due to the applied interpolation method is of the same order of magnitude as the one due to errors in the DEM. However, in the presented distributed model the step from a non-cAA grid cell to a cAA grid cell is dominated by the parameter with the greatest variability from one cell to its immediate neighbours. The mean absolute difference from a grid cell to its eight neighbours of the used DEM is 15.9 m, whereas the one of the temperature at 2,000 m a.s.l. is 0.0003 °C (corresponding to 0.04 m). This means that the altitude of the terrain is the decisive factor determining in the spatial distribution of the $rcELA_0$, superimposed by the precipitation field and the temperature field at 2,000 m a.s.l. over the Alps.

The non-linearity of the empirical relationship between precipitation and temperature at the ELA_0, Equation (1), is physically sound since, with increasing temperature, the rise of the ELA leads to a longer ablation season. Hence the influence of the temperature lapse rate becomes more dominant and the mass balance gradient increases. An additional amount of solid precipitation is needed to compensate the melting. Equation (1) includes an uncertainty of about 500 mm precipitation for a given temperature at the ELA_0. The corresponding confidence zone is a conservative approximation of the 95% confidence interval (Equation (2), Fig. 5). By applying this confidence zone in space, an uncertainty buffer with an average deviation of about ±190 m from the $rcELA_0$ of the reference run can be modelled, that accounts for measurement errors and for deviations of the $ltELA_0$ from the modelled $rcELA_0$ due to influences not included in the model (e.g., snow drift, avalanches, solar radiation). The general overestimation of the cAA on SE–SW slopes and underestimation on NE–NW slopes can be interpreted as differences, for example in solar radiation. Distributed mass balance models are needed (e.g., Klok and Oerlemans, 2002; Paul et al., in press; Machguth et al., subm.) in order to account for these topographic effects. The influence of the uncertainty in $rcELA_0$ on the cAA is indirectly proportional to the slope of the terrain, i.e., the steeper the terrain, the smaller the corresponding error of the cAA.

Beyond the range of precipitation and temperature values, used for the derivation of Equation (1), the uncertainties increase non-linearly. However, a comparison with worldwide data from Ohmura et al. (1992) shows that the empirical relationship derived from the 14 Alpine glaciers is also plausible for strongly continental and strongly maritime regions in the world. In Fig. 10, the 3-month summer temperature and the 6-month winter precipitation at the ELA_0 are plotted for the 14 Alpine glaciers in this study as well as for 69 glaciers worldwide as published in Ohmura et al. (1992). In addition, the exponential relationship derived only from the 14 Alpine glaciers is drawn, together with a shifted exponential curve that assumes a systematic precipitation measurement error of 23% (i.e., mean of the estimated measurement error of 15–30% for altitudes above 1,500 m a.s.l.; Sevruk, 1985). We have not combined these two data sets in order to derive a worldwide valid relationship between precipitation and temperature at the ELA_0, as they are not based on the same methodology, and in particular because the precipitation figures for Alpine glaciers existing in both data sets differ significantly. However, it is a future aim to extend the range of validity of the relationship, towards both more continental and more maritime regions. We recommend the determination of monthly, seasonal or at least annual precipitation and temperature at the ELA_0 of all glaciers with mass balance measurements. The determination of these two parameters together with the calculation of the ELA should become a standard for mass balance measurements.

The qualitative check of the cAA with the 1973 Swiss Glacier Inventory and with LandsatTM images dating from the 1990s, as well as the resulting AAR of 0.67 (calculated from the modelled cAA and the measured glacier area of the 1970s), confirm the overall accuracy of the modelled $rcELA_0$ and cAA. This can be quantitatively validated using a classification error matrix (Lillesand and Kiefer, 1994: 612–618) as soon as glacier outlines for the entire Alps are available and the real Alpine AAR value(s) are known. In addition, distributed mass balance values can be used to quantify the deviations of the $ltELA_0$ from the $rcELA_0$.

5.2. Findings from the distributed modelling in comparison with other studies

For the first time, $rcELA_0$ and cAA have been modelled at reasonable spatial resolution over the entire European Alps. Qualitative comparisons show that the modelled cAA of the reference model run (1971–1990) corresponds well overall to the real accumulation areas of Alpine glaciers. The cAA is clearly underestimated only between the Silvretta group and the Bernina group, in the eastern part of Switzerland, or only exists when the lower confidence boundary, Equation (3), is used instead of the reference run, based on Equation (1). In this region, the AIG is above many peaks, i.e. there is no cAA. The reason for this is found in the underestimation of the precipitation in that region. To decrease the AIG at the location of Silvretta Glacier to the terrain (so that the AIG becomes a $rcELA_0$), an increase in annual precipitation of 757 mm (+53%) would be needed. To sustain a glacier at that location another decrease in the $rcELA_0$ by 200 m is required, which would result in a total precipitation increase of about 1,800 mm (+125%). The probable cause of the underestimation of precipitation in the Silvretta region is the climatology by Frei and Schär (1998) and Schwarb et al. (2001). This is also supported by other studies. Machguth et al. (subm.) used the same precipitation data set as input for a distributed mass balance model applied in the Silvretta group. They concluded that annual precipitation in that region is underestimated by about 1,300 mm. In addition, the Hydrological Atlas of Austria (2003) differs in this part by about 1,600 mm from the precipitation climatology by Frei and Schär (1998) and Schwarb et al. (2001).

The sensitivity model runs show that a temperature change of ±1 °C would be compensated by a precipitation increase/decrease of 25%. The relative precipitation change corresponds to a mean absolute change over the greater Alpine region of about 300 mm. This is in agreement with Oerlemans (2001) and Braithwaite and Zhang (2000), who showed from mass balance modelling that a 25% increase in annual precipitation is typically needed to compensate for the mass loss due to a uniform warming of 1 °C. Kuhn (1981) calculated a rise of the ELA by 100 m if either the winter accumulation decreases by 400 mm or if the summer free air temperature increases by 0.8 °C. $MR_{T-0.6}$ leads to an average ELA_0-decrease of 75 m over the entire Alps and of 65 m within the 1973 outlines of the Swiss Glacier Inventory larger than 1 km^2. This corresponds well to the reconstructed ELA rise of 69 m between 1850 and 1973 presented by Maisch et al. (2000). The cAA difference of -26% between $MR_{T-0.6}$ and $MR_{1971–1990}$ is somewhat lower than the loss in glacier area (35%) between 1850 and the 1970s as estimated from inventories by Zemp et al. (in press). This suggests that either the model is not perfectly able to reproduce the area loss between 1850 and the 1970s, or that a rise in temperature by 0.6 °C cannot completely explain the corresponding glacier shrinkage. In the $MR_{T+3/P+10}$ the 10% increase in precipitation compensates for only a small amount of the ELA_0 rise caused by the temperature rise of 3 °C. The average rise of the $rcELA_0$ of 336 m clearly exceeds the upper boundary of the approximated confidence interval, based on Equation (4). As a consequence, many areas become ice-free. The cAA of the reference run is reduced by 74%. This relative area loss clearly indicates the dimension of such a change. These results correspond well with other studies that are based on different methods. Maisch et al. (2000) quote a glacier area loss in Switzerland of 75% for an ELA rise of 300 m. Haeberli and Hoelzle (1995) applied a parameterisation scheme to glacier inventory data to simulate potential climate-change effects on Alpine glaciers. They expect a reduction of glacier area and volume to a few percent of the values estimated for the 1850 glacierisation by the second half of the 21st century.

5.3. Potential applications

The presented approach is a valuable complement to current distributed mass balance models (e.g., Klok and Oerlemans, 2002; Paul et al., in press; Machguth et al., subm.). Since the distributed ELA_0-model needs only a minimum of input data, it is able to compute the $rcELA_0$ and the cAA over the entire Alps. Distributed mass balance models are (currently) not able to be applied to such large areas, but can be used for individual glaciers or within glacierised catchments to account for further important components of the energy balance (e.g., solar radiation, albedo, turbulent fluxes, mass balance-altitude feedback) and local, topographic effects (e.g., shading, avalanches, snow drift).

Equation (1) has the potential to estimate precipitation at glacier locations when the glacier ELA_0 and the temperature at the ELA_0 are known. The example of Silvretta Glacier demonstrates that the presented approach is a promising tool for estimating precipitation in high altitude terrain. However, in doing so, one has to take into account the uncertainties, both of the model and of the reconstruction of the ELA_0. In addition, Equation (1) seems to be plausible beyond the range of the parameters on which it is based (Fig. 10), but a future aim must be to enlarge this data set on an identical methodological basis. Here, availability and quality of the precipitation data at glacier ELA_0 will be the decisive element.

Regional climate models (RCM) have a horizontal resolution of about 20–50 km. These models use static glacier masks which indicate whether a specific grid cell is covered completely by ice or is totally ice-free. These glacier masks remain constant throughout the model simulation. Potential changes in the ice cover, their feedback to the atmosphere and their effects (e.g., enhanced summer runoff due to glacier melting) are not accounted for.

Current efforts are undertaken to represent mountain glaciers in RCMs on a subgrid scale (Kotlarski and Jacob, 2005.). Here, information on altitude-area distribution of glaciers for the initialisation of time-slice experiments and for model validation is urgently needed. Hence, quality and resolution of the approach presented here would definitely be good enough to provide an approximation of glacier altitude-area distribution for past and present states as well as for future climate scenarios.

6. Conclusions

Based on direct glaciological mass balance measurements for the period of 1971–1990, this study presents an empirical relationship between 6-month summer temperature and annual precipitation at the glacier ELA_0. Using GIS techniques and a DEM (SRTM3), this relationship is applied for the first time to a distributed modelling of the $rcELA_0$ and the cAA over the entire Alps at a spatial resolution of 3 arc seconds (approx. 100 m). It is shown that the altitude of the terrain is the decisive factor determining the spatial distribution of the $rcELA_0$, and hence also of the cAA, superimposed by the precipitation field and the temperature field at 2,000 m a.s.l. The empirical relationship is able to explain about 50% of the variance in precipitation at the glacier ELA_0 by the variance in temperature at that location. Hence, the uncertainties in the prediction of precipitation based on a given temperature and the ELA_0 at a single location are rather large. Nevertheless, it represents a promising tool for the estimation of precipitation in high altitude terrain based on a large number of ELA_0 values. The empirical relationship is used as a boundary condition for the distributed modelling of the $rcELA_0$. The mean absolute difference between modelled $rcELA_0$ and measured ELA_0 amounts to about 70 m. The uncertainties of the $rcELA_0$ are restricted to its location, and affect the cAA only in flat terrain. Under the assumption that local, topographic effects and components of the energy balance not included in the model do not change, the distributed modelling approach presented here is an adequate tool for studies of glacier sensitivity to changes in temperature and/or precipitation.

A sensitivity study shows that a summer temperature deviation of ±1 °C results in an average deviation of the $rcELA_0$ of +137/-125 m. An annual precipitation deviation of ±25% (an order of magnitude of 300 mm) leads to a deviation of the $rcELA_0$ of -114/+157 m. A summer temperature decrease of 0.6 °C lowers the ELA_0 by 75 m on the Alpine average and by 65 m within the outlines (> 1 km^2) of the 1973 Swiss Glacier Inventory. A summer temperature rise of 3 °C combined with an increase in annual precipitation of 10% results in an average rise in the $rcELA_0$ of 336 m and a reduction in the cAA in the order of 74%.

To enhance the statistical basis and to extend the range of validity of the empirical relationship, we recommend the determination of monthly, seasonal or at least annual precipitation and temperature at the ELA_0 of all glaciers with mass balance measurements. These two parameters together with the calculation of the ELA values should become a standard component of mass balance measurement series.

The presented approach is an excellent complement to distributed mass balance models. As the distributed $rcELA_0$-model requires only a minimum amount of input data to compute the $rcELA_0$ over the entire Alps, distributed mass balance models can then be used to account for further important components of the energy balance (e.g., solar radiation, albedo, turbulent fluxes, mass balance-altitude feedback) and local, topographic effects (e.g., shading, avalanches, snow drift) within individual catchments.

In conclusion, distributed modelling of the $rcELA_0$ can potentially contribute to the current efforts to include glacier altitude-area distribution of past, present and future glacier states in regional climate models.

Acknowledgements

We are indebted to the following institutes and organisations that placed their data at our disposal: Central Institute for Meteorology and Geodynamics in Vienna (temperature data), Institute for Atmospheric and Climate Science at the ETH in Zurich (precipitation data), Land Processes Distributed Active Archive Centre located at the U.S. Geological Survey in Sioux Falls (HYDRO1k), MeteoSwiss in Zurich (temperature data), National Aeronautics and Space Administration in Washington (SRTM3 and SRTM30), Swisstopo in Bern (DEM25L2), and World Glacier Monitoring Service in Zurich (mass balance measurements). Sincere thanks are given to the following colleagues for fruitful discussions and helpful comments in their research fields: Christoph Frei and Manfred Schwarb (precipitation and precipitation gradients in the European Alps), Sven Kotlarski (regional climate models), Frank Paul (DEM production and mass balance modelling), Max Maisch (Swiss glaciers), Nadine Salzman (climate modelling and interpolation techniques), Dominic Schuhmacher and the BABS group (statistical analysis). We gratefully acknowledge the constructive comments of Andrew G. Fountain, an anonymous referee and of the guest editor in charge Bruce Raup. Thanks go to Karen Hames and Susan Braun-Clarke for editing the English. The present study is funded mainly by the Swiss Federal Office of Education and Science (BBW-Contract 901.0498-2) within the EU programme ALP-IMP (Contract EVK2-CT-2002-00148).

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Tables

Table 1
Precipitation and temperature at the ELA_0 for the period of 1971–1990. ELA_0 is calculated from the linear relationship between ELA and the specific net balance. Annual precipitation (P_a, sum) and 6-month winter precipitation ($P_{O–M}$, sum) are derived from the Alpine precipitation climatology by Frei and Schär (1998) and Schwarb et al. (2001). Annual temperature (T_a, average), 6-month summer temperature ($T_{A–S}$, average) and 3-month summer temperature (T_{JJA}, average) are interpolated at 2,000 m a.s.l. from high altitude weather stations (see Fig. 1) and extrapolated to the ELA_0.

Glacier name	**LAT**	**LON**	**ELA_0**	**P_a**	**$P_{O–M}$**	**T_a**	**$T_{A–S}$**	**T_{JJA}**
	[°]	[°]	[m a.s.l.]	[mm]	[mm]	[°C]	[°C]	[°C]
Fontana Bianca (IT)	46.48	10.77	3177	1125	417	-5.43	-1.58	1.20
Gietro (CH)	46.00	7.38	3174	1250	604	-5.52	-1.63	1.21
Kesselwand (AT)	46.83	10.80	3106	1270	480	-5.19	-1.52	1.12
Careser (IT)	46.45	10.70	3088	1111	401	-4.86	-0.91	1.91
Vernagt (AT)	46.88	10.82	3076	1100	428	-5.03	-1.35	1.28
Gr. Aletsch (CH)	46.50	8.03	*2961	2926	1327	-4.18	-0.23	2.40
Hintereis (AT)	46.80	10.77	2933	1228	466	-4.13	-0.33	2.35
Rhone (CH)	46.62	8.40	2918	2031	1094	-4.04	-0.44	2.26
Gries (CH)	46.43	8.33	2852	2226	1036	-3.58	0.34	3.07
Plattalva (CH)	46.83	8.98	2772	1968	839	-3.14	0.64	3.34
Wurten (AT)	47.03	13.00	2770	2518	1063	-3.61	0.73	3.47
Silvretta (CH)	46.85	10.08	2755	1443	530	-2.93	0.89	3.56
Sonnblick (AT)	47.13	12.60	2731	2726	1135	-3.26	0.89	3.59
Limmern (CH)	46.82	8.98	2677	2012	857	-2.57	1.27	3.99

*The ELA_0 of Grosser Aletsch Glacier is calculated from mass balance data from the hydrological method, and ELA values interpolated from snow pits in the accumulation area of the glacier (Jungfraufirn) and aerial photographs.

Table 2
Overview of the seven model runs. Deviations in temperature and precipitation of the six sensitivity runs from the reference model run ($MR_{1971–1990}$).

T/P changes	**$MR_{1971–1990}$**	**MR_{T+1}**	**MR_{T-1}**	**MR_{P+25}**	**MR_{P-25}**	**$MR_{T-0.6}$**	**$MR_{T+3/P+10}$**
ΔT [°C]	0	+ 1	-1	0	0	-0.6	+3
ΔP [%]	0	0	0	+25	-25	0	+10

Table 3
Basin mean ELA_0 of the reference model run ($MR_{1971–1990}$) and ELA_0 deviations of the six sensitivity runs. Zonal borders come from the HYDRO1k DEM (http://LPDAAC.usgs.gov). In addition, zonal means within the 1973 outlines of the Swiss Glacier Inventory (Kääb et al., 2002; Paul et al., 2002) for glaciers larger than 1 km^2 are given. *The calculation of an $rcELA_0$ difference from the reference run is not possible, since for one model run the #AIG is above the highest peaks in of the corresponding hydrological basin.

Hydrological basin (no.)	$MR_{1971–1990}$ [m a.s.l.]	MR_{T+1} [m]	MR_{T-1} [m]	MR_{P+25} [m]	MR_{P-25} [m]	$MR_{T-0.6}$ [m]	$MR_{T+3/P+10}$ [m]
Lech (6985)	#	#	*	*	#	*	#
Isar (6988)	2592	+190	-137	-130	#	-60	#
Enns (7078)	2772	+131	-124	-110	+148	-78	+319
Traun (7080)	#	#	*	*	#	*	#
Linth (7124)	2784	+135	-134	-125	+149	-80	+340
Reuss (7130)	2784	+154	-131	-124	+167	-82	+366
Mur (7171)	2716	+130	-47	-47	+130	+18	#
Inn (7188)	3102	+118	-154	-142	+152	-95	+375
Aare (7194)	2795	+137	-125	-117	+150	-72	+310
Ill (7196)	#	#	*	*	#	#	#
Landquart (7210)	2807	+119	-80	-73	+113	-32	#
Ziller (7211)	2873	+150	-128	-121	+174	-83	+359
Arve (7237)	2895	+141	-120	-112	+170	-64	+360
Rhone (7246)	2966	+130	-132	-122	+147	-73	+315
Hinterrhein (7250)	2805	+126	-116	-103	+144	-71	+302
Drau (7261)	2810	+118	-96	-86	+145	-62	+309
Adige (7268)	3110	+135	-128	-114	+151	-80	+296
Vorderrhein (7272)	2896	+163	-44	-42	+168	-24	+327
Toce (7303)	2878	+161	-145	-134	+192	-92	+414
Adda (7332)	3144	+121	-194	-188	+171	-116	+351
Piave (7356)	2981	#	-83	-62	#	-66	#
Isere (7403)	3154	+131	-154	-142	+166	-90	+368
Aosta (7411)	3001	+112	-118	-107	+124	-65	+283
Drava (7420)	#	#	*	*	#	*	#
Chiese (7427)	3079	+137	-119	-110	+184	-91	#
F. Sesia (7438)	2980	+116	-117	-104	+113	-66	+295
Oreo (7487)	3159	+138	-135	-134	+165	-82	+325
D. Riparia (7538)	3144	+114	-138	-130	+190	-78	#
Durance (7557)	3183	+127	-193	-166	+169	-116	+344
Chisone (7581)	3148	+184	-83	-79	+184	-91	+368
Po basin (7605)	2987	#	-123	-108	#	-76	#
Po fount (7673)	#	#	*	*	#	*	#
Var (7737)	3077	#	-201	-172	#	-131	#
Argens (7741)	#	#	*	#	#	#	#
Liguria (7747)	#	#	*	*	#	*	#
All hydro basins	**2951**	**+137**	**-125**	**-114**	**+157**	**-75**	**+336**
Swiss glaciers	2946	+126	-118	-107	+140	-65	+308

Table 4
Slope-area distribution (in km^2) within the cAA (1971–1990) and within the 1973 outlines of the Swiss Glacier Inventory inside the cAA (1971–1990), computed in the Canton Valais, Switzerland (corresponding approximately to the Rhone basin). The glacier fraction of each slope class (in %) is calculated based on this distribution.

Data set	**0–10°**	**11–20°**	**21–30°**	**31–40°**	**41–50°**	**51–60°**	**61–70°**	**71–80°**	**81–90°**
cAA	220.0	370.7	290.3	171.7	70.8	13.4	1.5	< 0.1	0
Glacier area within cAA	196.3	288.7	169.8	67.7	22.8	3.8	0.4	0	0
Glacier fraction	89	78	58	39	32	29	29	0	0

Figures

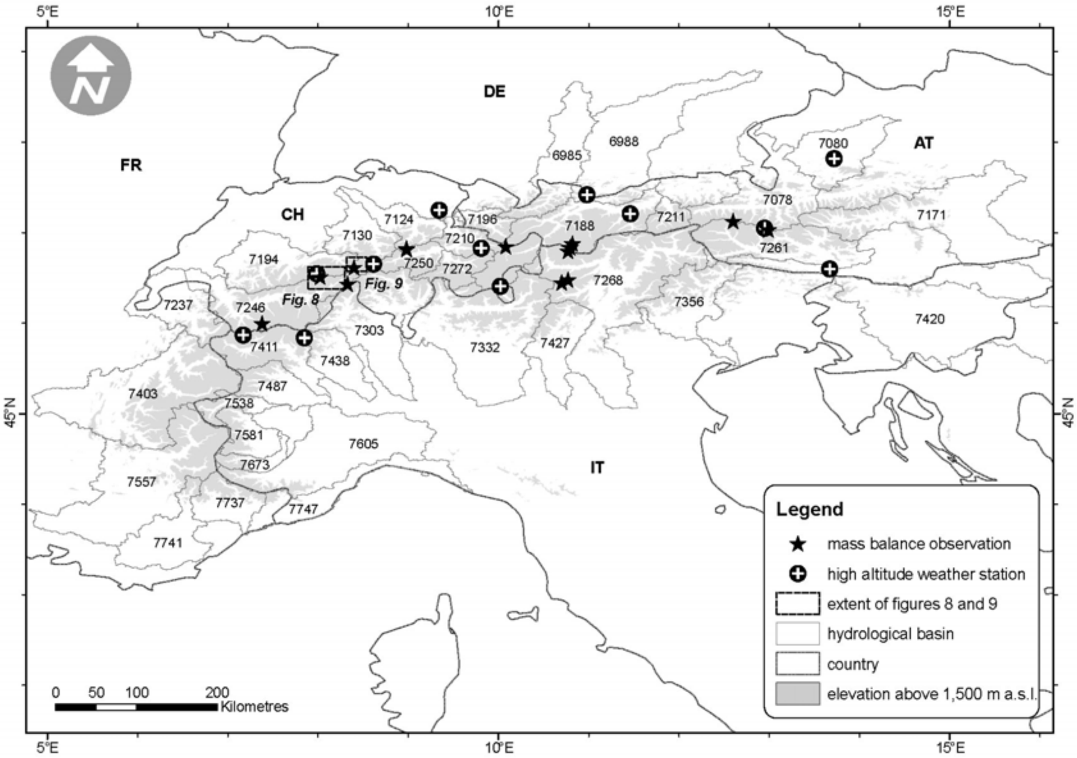

Fig. 1. Alpine mountain ridge (shaded in grey) with available mass balance observations (glaciers from west to east: Giétro, Grosser Aletsch, Gries, Rhone, Limmern, Plattalva, Silvretta, Caresèr, Fontana Bianca, Hintereis, Kesselwand, Vernagt, Sonnblick and Wurten) and high altitude weather stations (from west to east: Gr. St. Bernhard, Lago Gabiet, Junfraujoch, Gütsch ob Andermatt, Säntis, Weissfluhjoch, Bernina Hospiz, Zugspitze, Patscherkofel, Sonnblick, Villacher Alpe and Feuerkogel). Hydrological basins from the HYDRO1k DEM (http://LPDAAC.usgs.gov) are labelled with the corresponding basin IDs. The spatial extent of Figures 8 and 9 is indicated by dashed black lines.

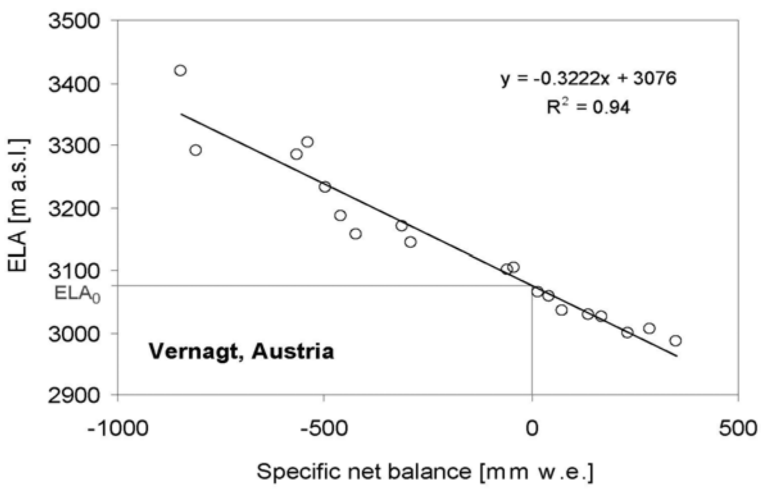

Fig. 2. Linear relationship between ELA and specific net balance of Vernagt Glacier (Austria) for the period 1971–1990. The ELA_0 of 3,076 m a.s.l. represents the calculated ELA value for a zero net balance, i.e., a hypothetical steady state.

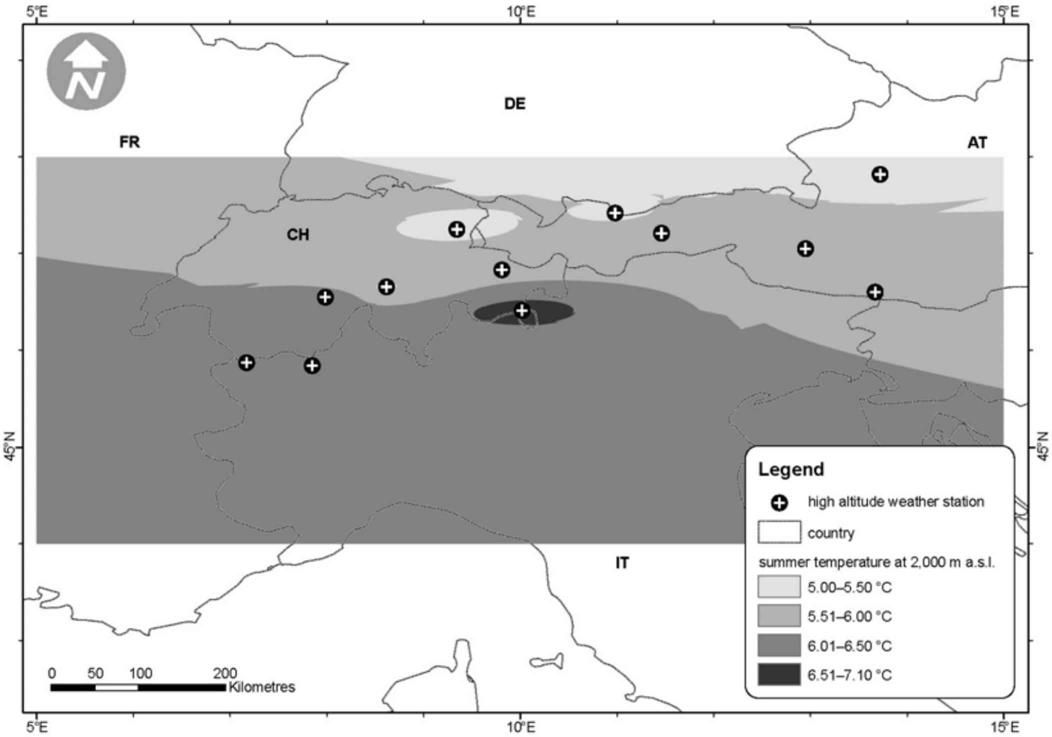

Fig. 3. The 6-month summer temperature field at 2,000 m a.s.l. (1971–1990) as interpolated from 12 high altitude weather stations. A seasonal lapse rate of 0.67 °C per 100 m was used to altitude-adjust the seasonal mean temperature at weather stations to 2,000 m a.s.l. The IDW (Watson and Philip, 1985) method was used for the interpolation.

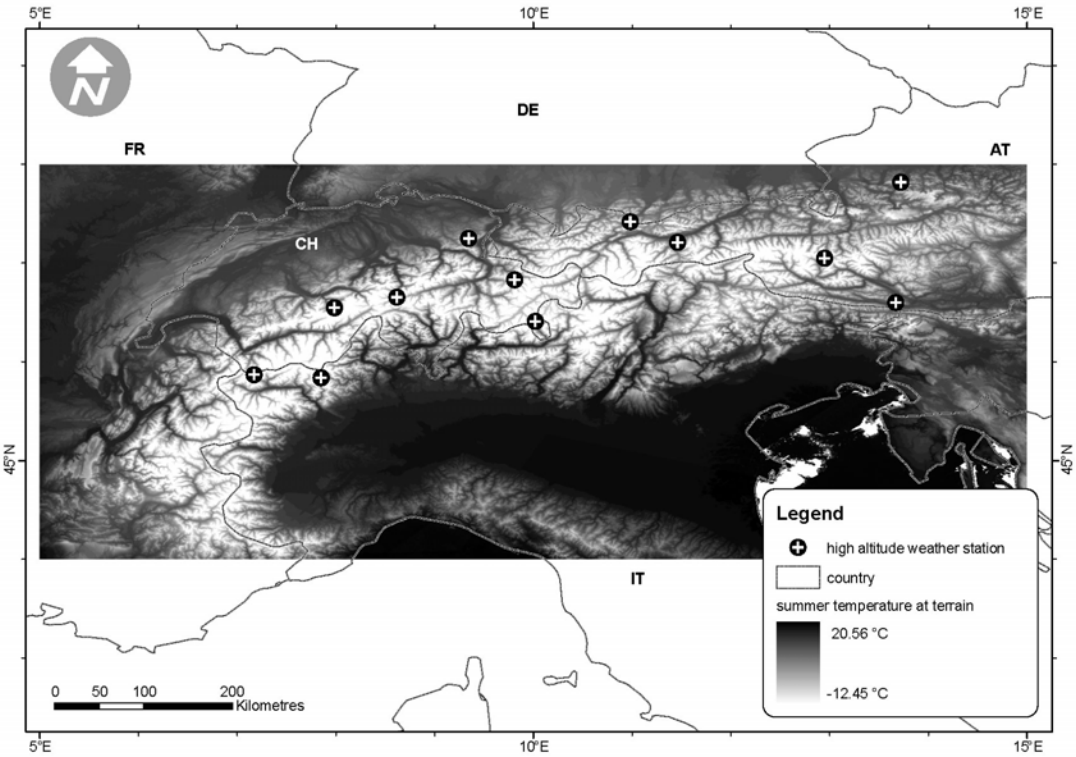

Fig. 4. The 6-month summer temperature on terrain (1971–1990). The data set was produced by extrapolating the temperature field at 2,000 m a.s.l. (Fig. 3) to the altitude of the terrain (SRTM3, competed with SRTM30) using a seasonal lapse rate of 0.67 °C per 100 m.

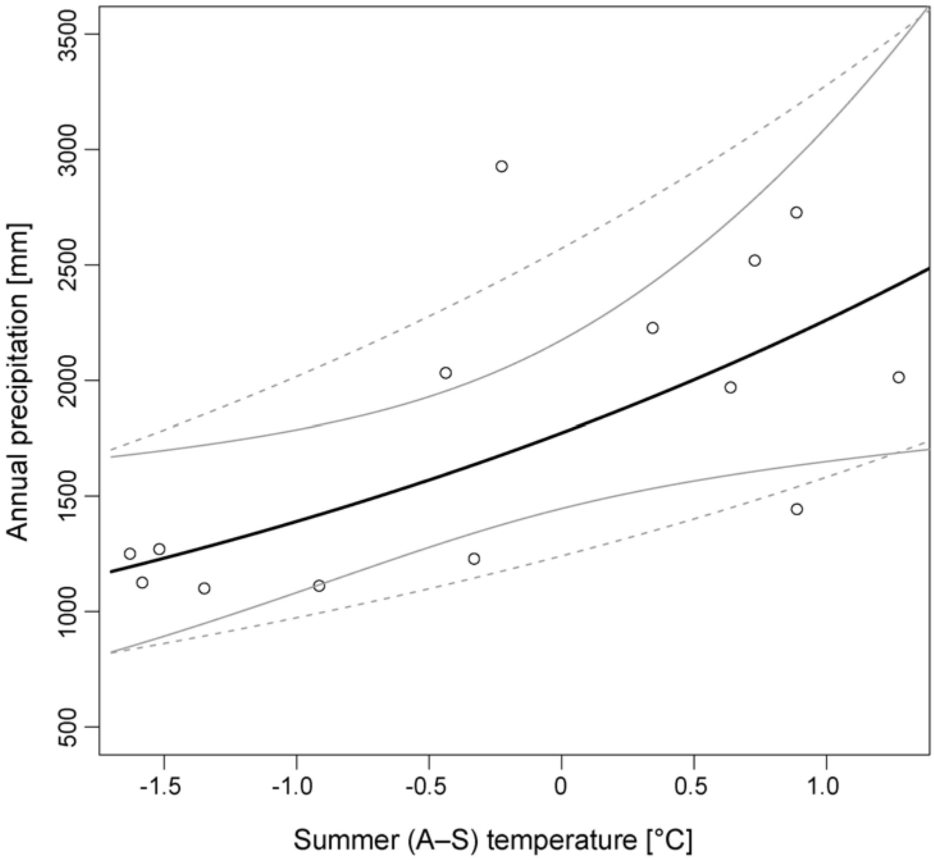

Fig. 5. Annual precipitation and 6-month summer temperature at the ELA_0 for the period 1971–1990 (cf. Table 1). The exponential regression model (Equation (1); black line) performs with a R^2 of 0.49. The boundaries of the 95% confidence interval (Equation (2); grey line) are conservatively approximated by shifting the regression model (Equations (3) and (4); dashed grey lines).

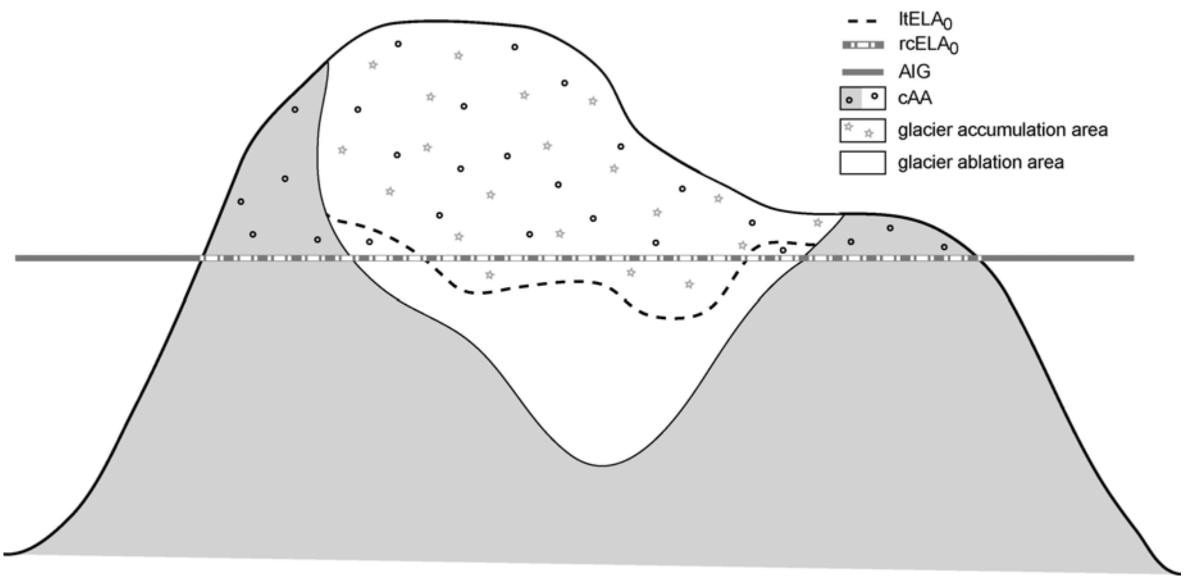

Fig. 6. Diagram to explain the concept of the AIG and its association with $ltELA_0$, $rcELA_0$ and cAA. Figure modified after Lie et al. (2003a).

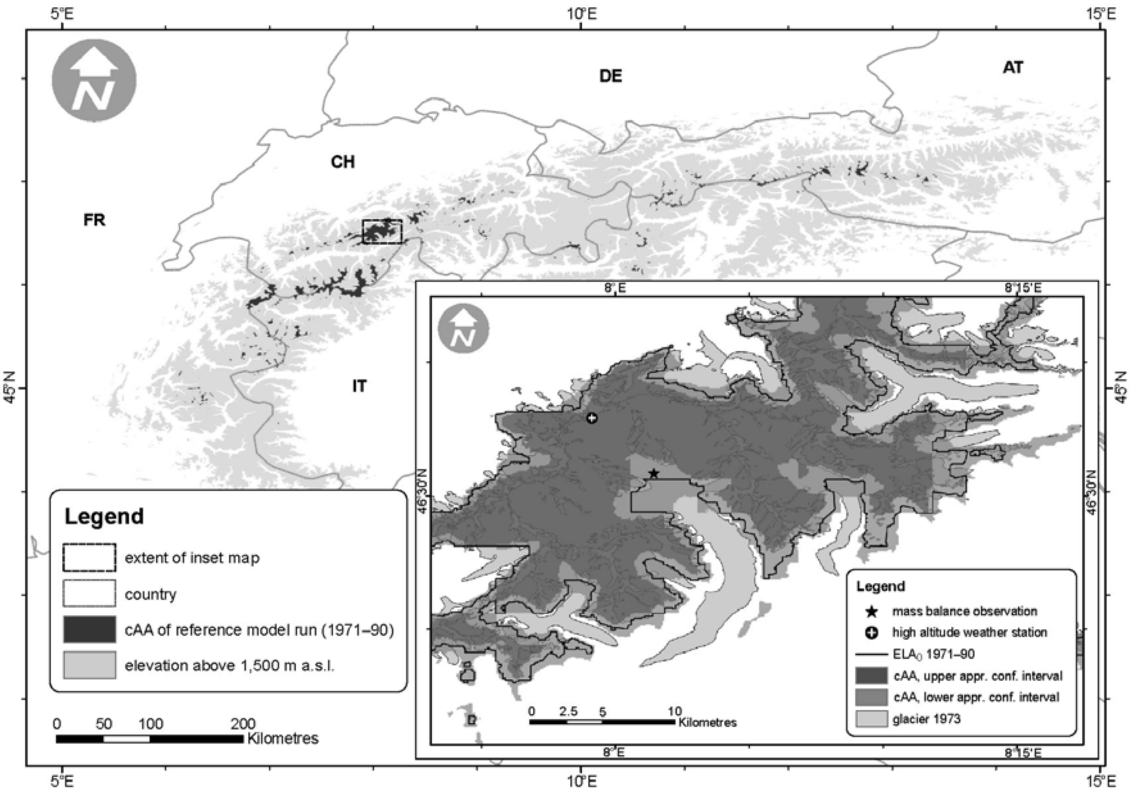

Fig. 7. cAA of the reference model run (1971–1990). The inset map shows the ELA_0 of the reference model run (black line) and the cAAs of the corresponding lower (grey) and upper (dark grey) boundaries of the approximated confidence interval. Glacier extents of 1973 from the Swiss Glacier Inventory are shown in light grey.

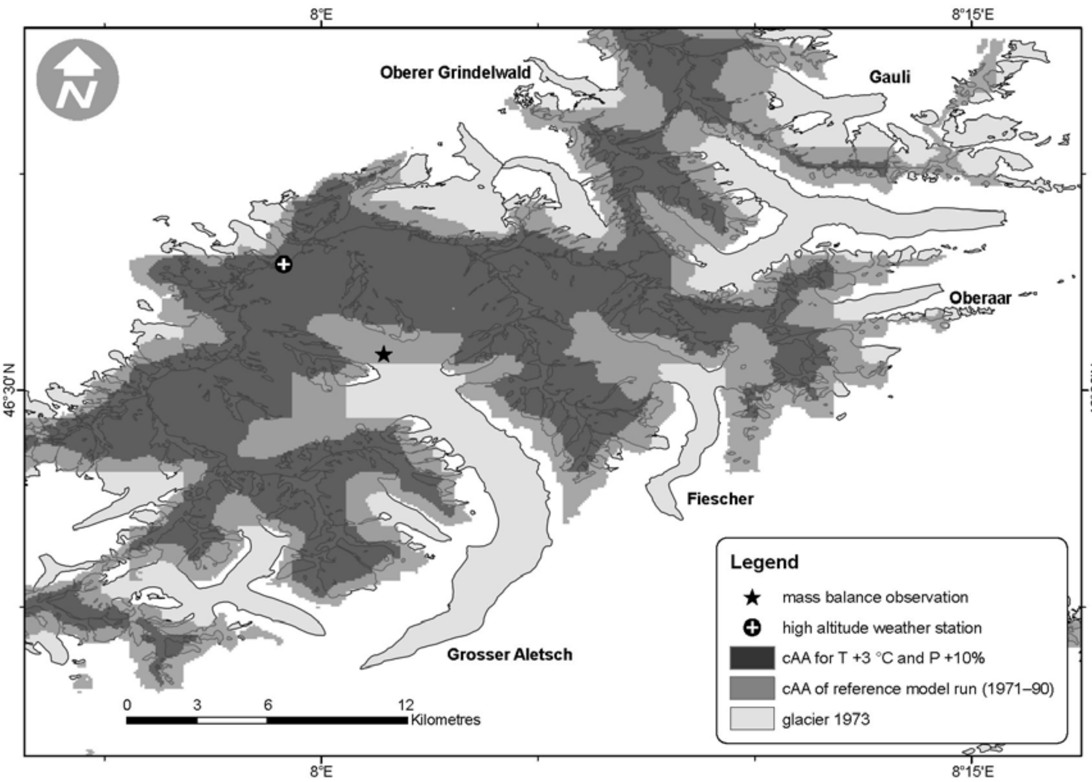

Fig. 8. cAA of the reference model run (1971–1990, grey) and cAA of the $MR_{T+3/P+10}$ (dark grey) in the region of Grosser Aletsch glacier, Switzerland. Glacier extents in 1973 from the Swiss Glacier Inventory are shown in light grey.

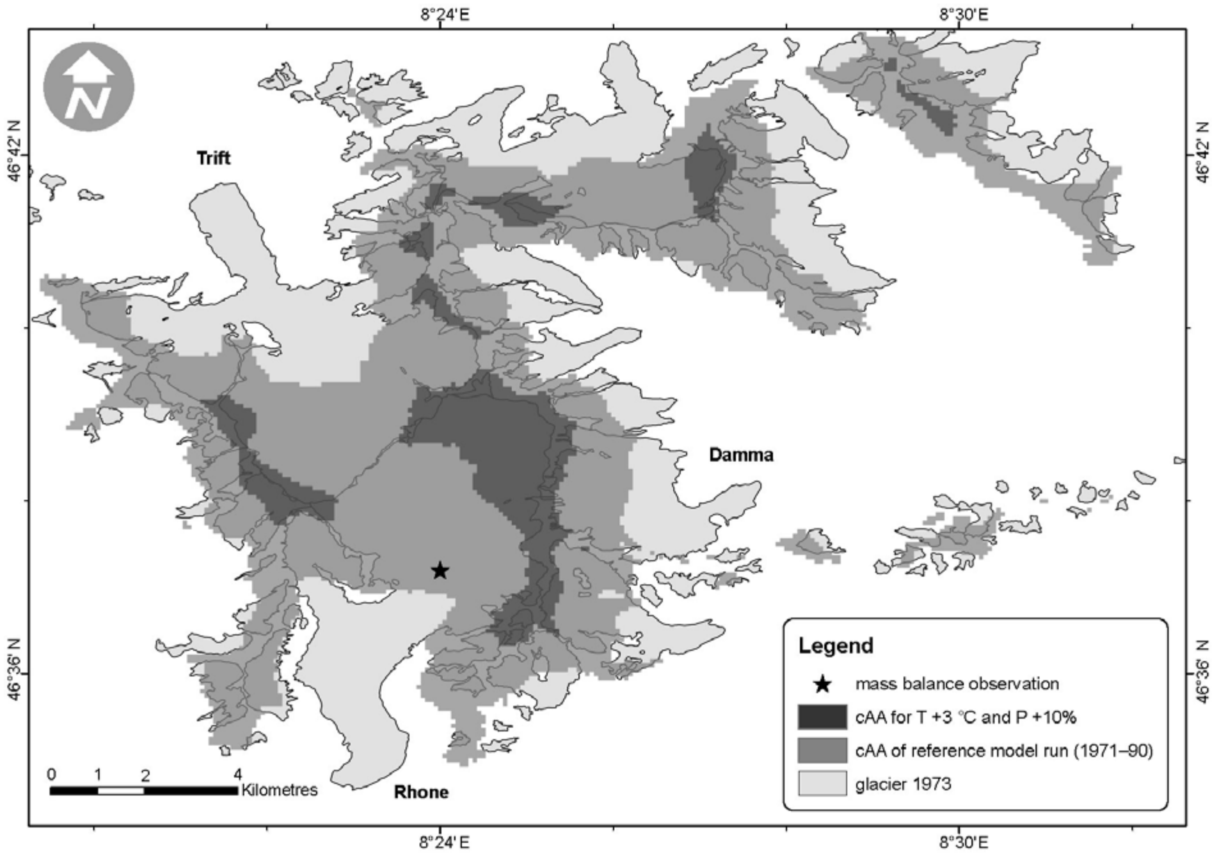

Fig. 9. cAA of the reference model run (1971–1990, grey) and cAA of the $MR_{T+3/P+10}$ (dark grey) in the region of Rhone glacier, Switzerland. Glacier extents in 1973 from the Swiss Glacier Inventory are shown in light grey.

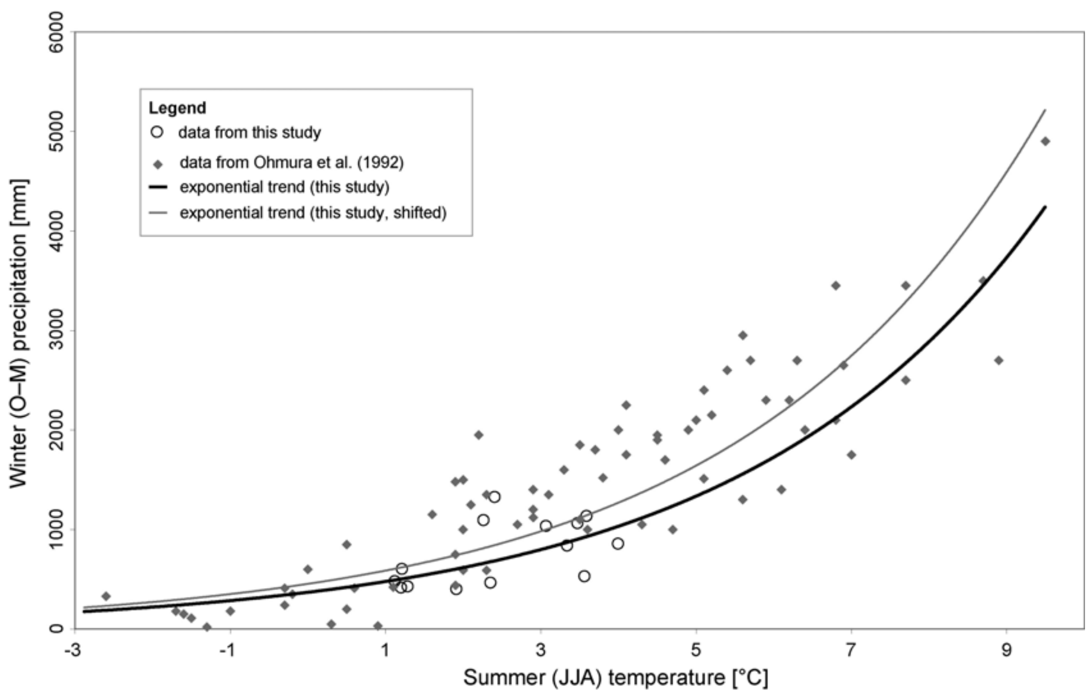

Fig. 10. The 6-month winter precipitation and the 3-month summer temperature at the ELA_0. Comparison between the data set presented in this study (black circles) and the one by Ohmura et al. (1992; grey diamonds). The exponential model (black line) based on the 12 data points from this study is shown together with a shifted exponential curve assuming a systematic precipitation measurement error of 23% (grey line).

Paper V

Alpine glaciers to disappear within decades?

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[1] Glacier cover in the entire European Alps has been computed for different climate-change scenarios combining digital terrain information with satellite-derived glacier changes and a model for equilibrium line altitudes. A 3 °C warming of summer air temperature would reduce the currently existing Alpine glacier cover by some 80%, or up to 10% of the glacier extent of 1850. In the event of a 5 °C temperature increase, the Alps would become almost completely ice-free. Annual precipitation changes of ±20% would modify such estimated percentages of remaining ice by a factor of less than two. The evolution of the Alpine glacierization will present a crucial opportunity for a quantitative verification of climate scenarios.

1. Introduction

[2] Impacts on cold mountain ranges from ongoing and potentially accelerating climate change are especially pronounced in regions above the timberline where effects related to perennial surface ice reflect increasing atmosphere/earth energy fluxes with extraordinary clarity [*Royal Swedish Academy of Sciences*, 2002]. Many mountain ranges have lost a significant proportion of their glacierization during the past 150 years with strong acceleration occurring in the past two decades [e.g., *Haeberli et al.*, 2005a, 2005b]. The shrinking of mountain glaciers is indeed the most obvious indication in nature of fast if not accelerating climate change on a worldwide scale. The predicted global temperature increase [*Houghton et al.*, 2001] is likely to induce dramatic scenarios of future glacier developments including complete deglaciation of entire mountain ranges. Such future scenarios of glacier vanishing have thus far not been assessed quantitatively from spatial climatologies on an Alpine-wide scale, but are likely to affect landscape appearance, slope stability, the water cycle, sediment loads in rivers and natural hazards far beyond the range of historical and Holocene variability [*Watson and Haeberli*, 2004; *Barnett et al.*, 2005].

[3] In this study we apply an integrated approach, combining in-situ measurements, remote sensing techniques and numerical modeling to the European Alps. These techniques allow to quantitatively assess past as well as potential evolutions of area and volume of a glacier ensemble within an entire mountain chain. Glacier cover in the entire European Alps has been computed for different climate-change scenarios using satellite-derived glacier changes and a digital terrain model (DTM) together with a distributed model for equilibrium line altitudes (ELA). We thereby demonstrate the possibility of fast glacier disappearance within the European Alps, as well as the potential of new technologies (i.e. satellite imagery, DTM, geo-informatics) to use information from glacier monitoring in mountain regions for quantification of global climate-change scenarios (Figure AM1[1]).

2. Glacier fluctuations from 1850–2000

[4] Information on glacier fluctuations in the European Alps is available from earlier and recent glacier inventories [*Haeberli et al.*, 1989; *Maisch et al.*, 2000; *Kääb et al.*, 2002; *Paul et al.*, 2002; Figure AM2] together with data compilations on past glacier fluctuations [*Zemp et al.*, in press b; Figure AM2] National glacier inventories in the 1970s yield a total glacier

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[1] Auxiliary material (AM) is available via Web browser or via Anonymous FTP from ftp://ftp.agu.org/apend/" (Username = "anonymous", Password = "guest").

area of 2909 km^2 [*Haeberli et al.*, 1989]. During the mid-1970s, glacier mass balances were close to zero or slightly positive (Figure AM3), many shorter glacier tongues slightly re-advanced and, hence, most glaciers were probably quite close to equilibrium conditions. The fact that the time basis for the corresponding inventory data is not uniform (Austria 1969, France 1967–71, Germany 1975, Italy 1975–84 and Switzerland 1973, cf. [*Zemp et al.*, in press b]), therefore, plays a minor role: the center point of the corresponding time interval is thus defined as 1975. Detailed reconstructions of glacier areas around AD 1850 – the maximum extent for most glaciers in the European Alps at the end of the Little Ice Age – are available for the Swiss [*Maisch et al.*, 2000] and Austrian Alps (unpublished). The latest glacier inventory data based on satellite images is again available for most of the Swiss Alps in 1998/99 [*Kääb et al.*, 2002; *Paul et al.*, 2002], hereafter attributed to the year 2000 for the sake of simplicity. The Alpine glacier area in 1850 and 2000 is extrapolated by applying relative area changes for individual glacier size classes from the Swiss Alps to the corresponding entire Alpine glacier sample from 1975 (Table AM1). This extrapolation reveals an overall loss in Alpine glacier area of 35% from 1850 up until 1975 (-2.8% per decade) and almost 50% by 2000 (-3.3% per decade). The area reduction between 1975 and 2000 is about 22% (-8.8% per decade), mainly occurring after 1985 (i.e., -14.5% per decade) as glacier fluctuation measurements and satellite-derived data have clearly shown [*Paul et al.*, 2004; *Zemp et al.*, in press b; Figure AM3]. Disintegration and 'down-wasting' have been predominant processes of glacier decline during the most recent past [*Paul et al.*, 2004].

3. Assessment of glacier area loss during the 21st century

[5] Potential future area changes for the entire Alps are estimated by two independent methods. The first method is a purely empirical one that relates documented rates of area change for altitudinal bands (Figure AM4) to scenarios of glacier shrinking, ranging from 'continued loss' (area reduction for the period 1850–1975), 'accelerated loss' (loss from 1975–2000), 'strongly accelerated loss' (period 1985–2000) and 'extreme loss' (using a doubled 1985–2000 loss rate). These scenarios cover the range of documented glacier shrinking rates and are related to a 20th-century warming of about 1 °C in the European Alps [*Böhm et al.*, 2001]. The scenarios of future area losses (Figure 1) illustrate that the scenario of 'accelerated loss' would drastically reduce Alpine glacier areas within this century and that the scenario of extreme ice loss would cause most of the presently existing glaciers in the Alps to disappear within decades as large parts of the ice is located below 3000 m a.s.l. The 'extreme loss' scenario should be seen as an upper limit assumption but may not be unrealistic: hot-dry conditions like in the summer of 2003, which could occur at shorter and shorter intervals [*Schär et al.*, 2004] and involve strong reinforcing effects (albedo feedback, mass balance/altitude feedback, glacier down-wasting and collapse) could indeed soon bring about such a situation.

[6] Glacier health is primarily influenced by air temperature, while precipitation is the second most important climatic factor affecting their condition [*Kuhn*, 1981; *Oerlemans*, 2001]. The ELA on a glacier is a theoretical line which defines the altitude at which annual accumulation equals ablation. It represents the lowest boundary of climatic glacierization – that is, where the glacierization can begin. Hence, the second approach is a statistically calibrated and distributed model of ELA after *Lie et al.* [2003] that utilizes an empirical relation between 6-month summer air temperature ($T_{A\text{–}S}$, in degree Celsius) and annual precipitation (P_a, in millimeter) at the steady-state ELA (ELA_0) [*Zemp et al.*, accepted]:

$$P_a = 1773 \cdot e^{0.2429 \cdot T_{A\text{–}S}}$$

[7] The relation is obtained from long-term mass balance data from 14 Alpine glaciers [*Haeberli et al.*, 2005a] in combination with gridded precipitation [*Frei et al.*, 1998; *Schwarb*

et al., 2001] and temperature (interpolated from twelve high-altitude weather stations, cf. [*Zemp et al.*, accepted]) climatologies and a DTM of 100 m cell size, resampled from the DTM of the Shuttle Radar Topography Mission [*cf., Rabus et al.*, 2003]. Its application to the entire European Alps enabled distributed modeling of the regional climatic ELA for zero mass balance ($rcELA_0$) and the corresponding climatic accumulation area (cAA) above it (see Figure AM5 for a diagram explaining this concept and *Zemp et al.* [accepted] for methodological details). The accumulation area over the entire Alps obtained from the reference model run is 1950 km^2 and agrees well with accumulation areas of mapped glaciers (Figure AM6).

[8] The impact on glacier areas as related to scenarios of temperature and precipitation change is illustrated in Figure 2. Atmospheric warming of 3 °C in summer (April to September) accompanied by an increase of 10% in annual precipitation would, for instance, raise the $rcELA_0$ by 340 m and reduce the cAA by 75% compared to the 1971–1990 reference period. Depending on the climate scenario chosen, this could take place towards the middle or the end of the century [*Houghton et al.*, 2001]. Due to the strong warming in the past two decades, more than one-third of this glacier area reduction has already been taking place [*Paul et al.*, 2004; *Zemp et al.*, in press b]. An increase in summer air temperature of 5 °C would reduce the glacier cover by more than 90% as compared to the reference period. Precipitation changes of ±20% would modify such estimated percentages of remaining ice by a factor of less than two. Many individual mountain ranges within the Alps would become ice-free under such conditions and only rather small glacier remnants would persist in a few regions with the highest mountain peaks (Figures 3 and AM6). These results confirm earlier (independent) estimates from modeling of statistical glacier data [*Maisch et al.*, 2000; *Haeberli and Hoelzle*, 1995] and shows that the calculation is robust.

4. Estimations of past, present and future ice volumes

[9] Changes in glacier volume are calculated by multiplying representative mass balance values with the average surface area (the mean of the areas at the beginning and end) of a given time period. Mean mass balance of nine Alpine glaciers between 1975 and 2000 was almost -0.5 m water equivalent (w.e.) per year (Figure AM3). This is about twice the loss rate reconstructed from cumulative length change for the time period after 1850 [*Haeberli and Hoelzle*, 1995; *Hoelzle et al.*, 2003; *Steiner et al.*, 2005] and characteristic long-term mass changes during the past 2000 years [*Haeberli and Holzhauser*, 2003]. The cumulative balance of -12 m w.e. over a mean glacier area of 2590 km^2 during the same time interval (1975–2000) indicates a lower limit of the corresponding volume loss of 30 km^3. As average slope and ELA have increased, but glacier size (as well as altitudinal extent, mass flux and driving stress) decreased, the percentage of volume loss must be even greater than the calculated area loss of 22% (cf. remarks on volume estimations in auxiliary material). Corresponding estimates (25–30% volume loss) show that the glaciers in the Alps have lost an average of 1% of their volume per year since 1975. On the same basis, total Alpine ice volumes can be estimated roughly as 105 ±15 km^3 in 1975, and 75 ±10 km^3 at the turn of the century [*Paul et al.*, 2004], i.e., considerably lower than the 130/100 km^3 estimated earlier [*Haeberli and Hoelzle*, 1995]. Total glacier volume for the end of the Little Ice Age (around 1850) with an extrapolated total glacierized area of 4475 km^2 is estimated at some 200 km^3 or more, and is now close to one-third of this value.

[10] Mean ice depth over the entire remaining glacier area of the Alps, calculated as the quotient of ice volume and area in 2000, is only about 30–35 meters. The average mass balance of -2.5 m w.e. in the extreme year 2003, therefore, eliminated an estimated 8% of the remaining Alpine ice volume within one single year [*Haeberli et al.*, 2005a; *Zemp et al.*, in press a]. The following year 2004 with an average mass balance of -1 m w.e. reduced an additional 3%, leading to about 10% volume loss in only two years. Extremely hot and dry

summers such as 2003 thus not only induce strong positive feedbacks, but also eliminate increasing percentages of shrinking total ice volume. It is likely that five rather than ten repetitions within the coming decades of conditions as in 2003, would bring out this scenario of widely deglaciated Alps. As 90% of all Alpine glaciers are smaller than 1 km^2, the probability that most glaciers in the European Alps will disappear within the coming decades (Figure 3) is indeed not insignificant. The few largest valley glaciers with maximum ice thicknesses of several hundred meters will be able to resist such warming effect for somewhat longer. However, reinforcing mechanisms such as the mass balance/altitude feedback or the development of glacier lakes will also increasingly enhance their wasting.

5. Discussion and conclusions

[11] The major sources of error and corresponding consequences for the three methods (a: scenarios for changes in glacier hypsography, b: distributed ELA model, c: volume estimations) are: for a), the representativity of the analyzed sub samples for the entire Alpine glacierization, which might bias the rates of ice loss due to regional peculiarities, for b), the neglect of topographic effects (e.g., snow drift, avalanches, radiation) leading to differences between the modeled $rcELA_0$ and the local topographic ELA_0, and the assumption of a constant accumulation area ratio, resulting in an uncertainty in the extrapolation of glacier changes from the modeled cAA to the total glacier area, and for c) the representativity of average reconstructed/measured mass balance values from the used glacier samples for the entire Alpine glacierization in 1850/1975/2000, resulting in estimated thickness changes biased by mid-size glaciers [cf. *Zemp et al.*, in press b].

[12] Modern strategies of glacier observations established within international monitoring programs make use of fast-developing new technologies and relate them to traditional approaches in order to apply integrated, multilevel concepts [*Haeberli*, 2004]. The combination of in-situ measurements with remote sensing, digital terrain information and numerical models thereby allows for comprehensive views by assimilating individual observational components (length, area, volume/mass change) over different scales in space (local, regional, global) and time (past and future, years to decades and centuries). The resulting information and comparison with model applications for entire mountain ranges based on this do indeed represent a decisive possibility for verifying climate models and future scenarios of global climate change. Glacier changes as a function of climate change are not only easily observed, they are also comprehensible, in their basic physical principles, to a large public. This makes them unique demonstration objects and key variables for early-detection strategies in global climate-related observations [*World Meteorological Organization*, 2003]. By simply looking at the evolution of glaciers in the mountain ranges of the world, coming generations will be able to define and to physically see which scenario of climate change has taken place.

[13] Acknowledgements. We are indebted to the following institutes and organizations that placed their data at our disposal: Central Institute for Meteorology and Geodynamics in Vienna (temperature data), Institute for Atmospheric and Climate Science at the ETH in Zurich (precipitation data), MeteoSwiss in Zurich (temperature data), National Aeronautics and Space Administration in Washington (SRTM3 and SRTM30), Swisstopo (DEM25), National Point of Contact (satellite data) and World Glacier Monitoring Service in Zurich (glacier data). We thank all national correspondents, principal investigators and contributing institutes who have been providing the World Glacier Monitoring Service with data for many years! Thanks go to Susan Braun-Clarke for editing the English. This study is funded mainly by the Swiss Federal Office of Education and Science (BBW-Contract 901.0498-2) within the EU program ALP-IMP (Contract EVK2-CT-2002-00148).

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Figures

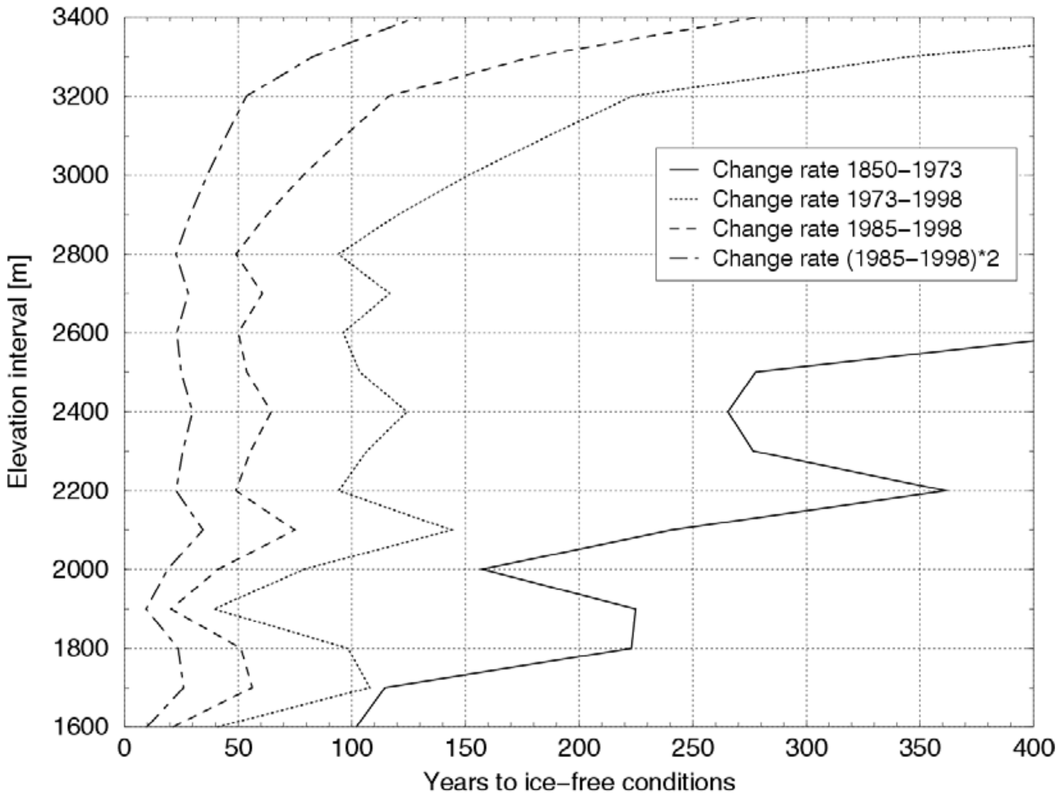

Figure 1. Years to ice-free elevation bands as obtained from the observed change in hypsography for two glacier samples (1850–1973 and 1973–1998, Figure AM4). The glacier covered area in 1973 (1998) has been divided by the respective area loss per elevation band and multiplied with the number of years in the respective period (125, 25, 13, 6). Compared to the 1850–1973 period, there is not really a dependence of the change on elevation up to 2800 m a.s.l. for the more recent period.

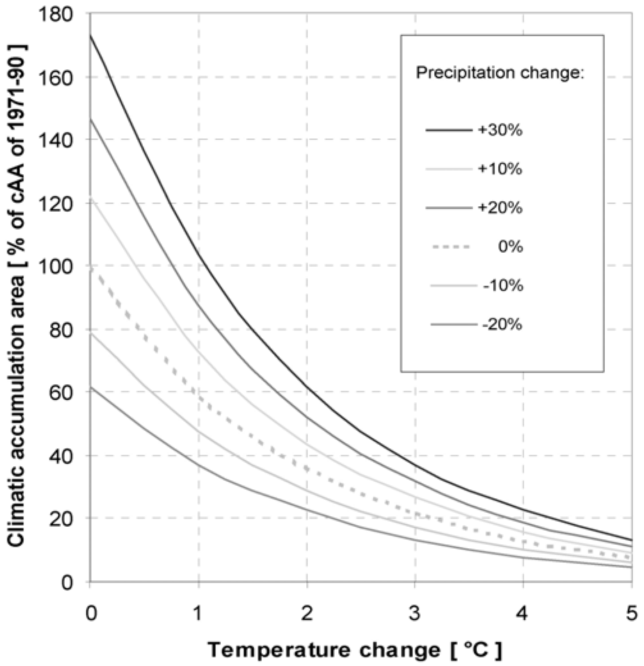

Figure 2. Modeled climatic accumulation area (cAA) according to changes in 6-month summer temperature and/or annual precipitation. The total of 100% refers to the cAA of the reference model run (1971–1990) and amounts to 1950 km^2. The changes in temperature and precipitation cover the range of the IPCC-scenarios [*Houghton et al.*, 2001] for the European Alps by the end of the 21st century. The dotted line refers to pure summer temperature changes from 0 to +5 °C. The other lines represent combined temperature and precipitation changes (order of the lines correspond to legend).

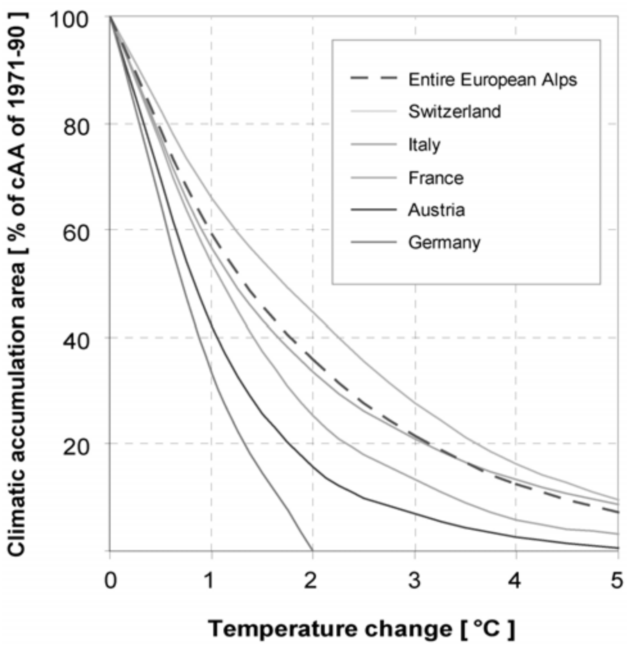

Figure 3. Modeled remains of Alpine glacierization as a consequence of a 1–5 °C warming of the 6-month summer temperature. The total of 100% refers to the cAA of the reference model run (1971–1990) and amounts to 1950 km^2 for the entire European Alps. The 100%-marks of the other lines refer to the fraction of glacierization of the corresponding Alpine country (order of the lines correspond to legend).

Auxiliary Material Submission for Paper

Alpine glaciers to disappear within decades?
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Introduction:
This auxiliary material contains six figures, one animation, one table and a text supporting the results of the present study. Glacier inventory and fluctuation data are provided by and available from World Glacier Monitoring Service (www.wgms.ch).

References used in auxiliary material:
Haeberli, W., H. Bösch, K. Scherler, G. Østrem, and C. C, Wallén (Eds.) (1989), *World Glacier Inventory – Status 1988*, 458 pp., IAHS(ICSI)/UNEP/UNESCO, Nairobi.

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Paul, F., A. Kääb, M. Maisch, T. W. Kellenberger, and W. Haeberli.(2002), The new remote-sensing-derived Swiss Glacier Inventory: I. Methods, *Ann. Glaciol., 34*, 355–361.

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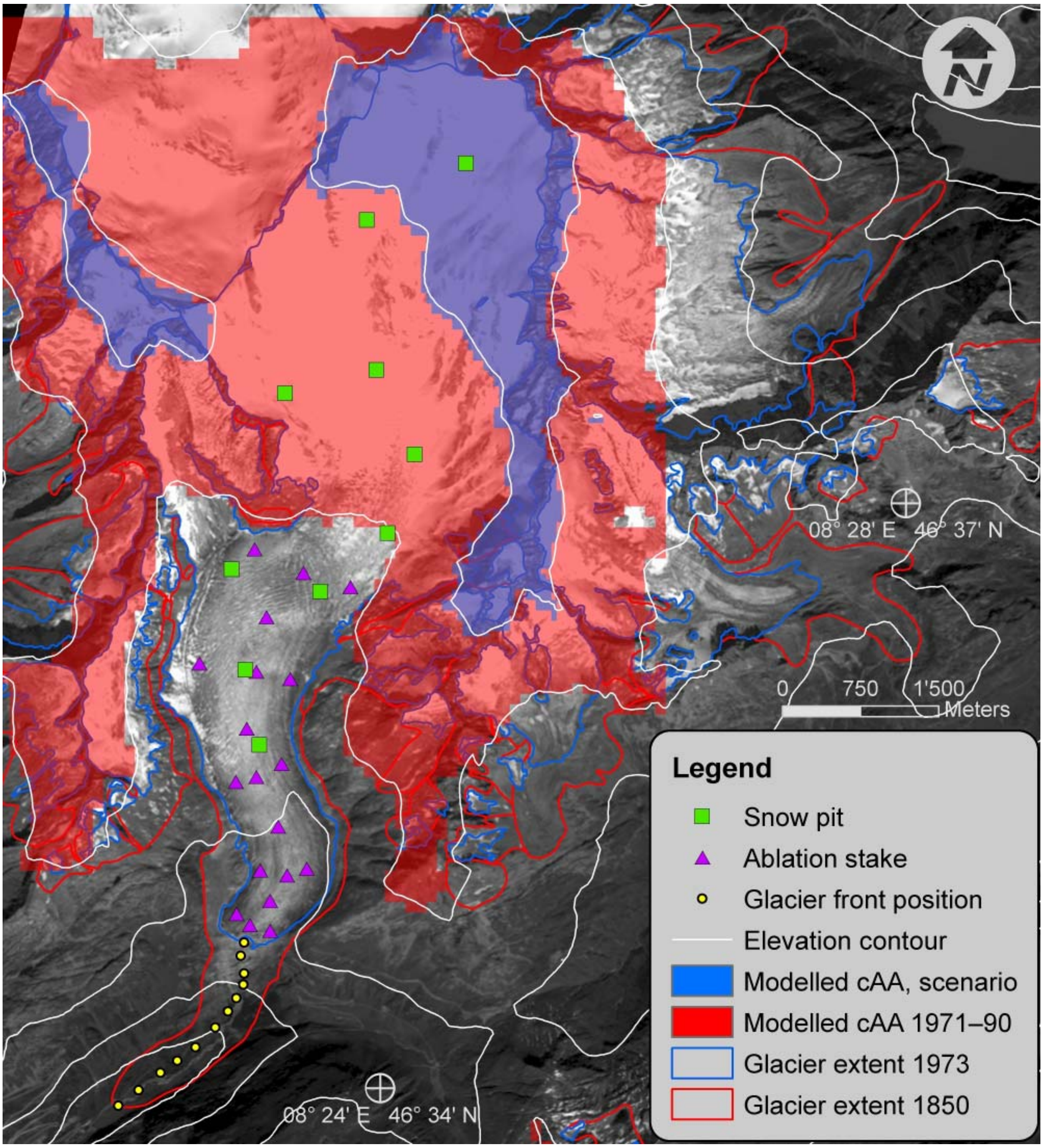

Figure AM1. The fundamental elements of the presented approach, following international monitoring strategies [*Haeberli*, 2004], are exemplified with a SPOT-2 panchromatic satellite image (17. Sep.1992) of the Rhone Glacier, Switzerland: in-situ observations (glacier front positions, locations of stakes and pits from mass balance measurements) are combined with glacier outlines derived from topographic maps and satellite images and supplemented with the climatic accumulation areas (cAA) of a reference period (1971–1990) and of a future scenario (T_{A-S} +3 °C, P_a +10%), modeled from gridded climatologies and digital terrain models (DTM). Elevations of the Rhone glacier range from about 1750 m a.s.l. at the tongue of the 1850-extent to about 3550 m a.s.l. in the accumulation area. Locations of in-situ observations are idealized but based on existing measurements. For the geographical location of the figure extent, see Figure AM2, rectangle B.

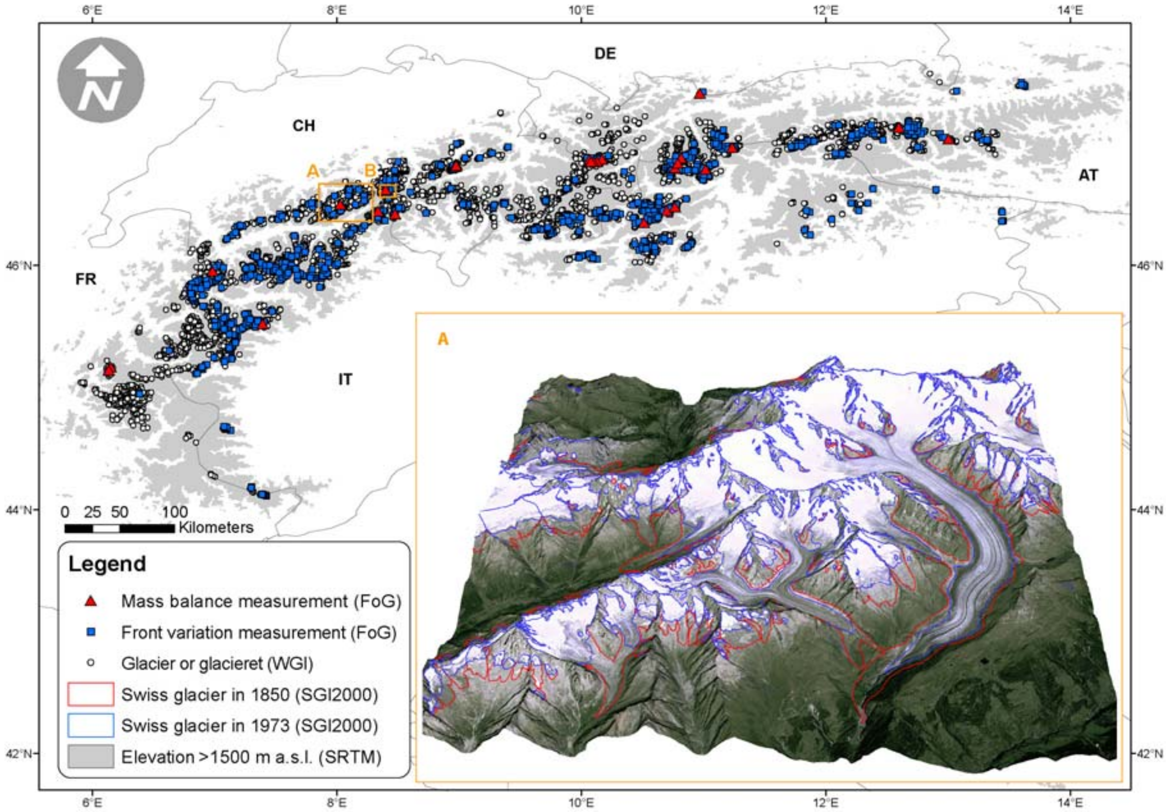

Figure AM2. Geographical distribution of available glacier information in the European Alps: static information on all Alpine glaciers and glacierets [*Haeberli et al.*, 1989], glacier front variation and mass balance measurements [*Haeberli et al.*, 2005a, b] as well as digital outlines of glaciers from the Swiss Glacier Inventory [*Kääb et al.*, 2002; *Paul et al.*, 2002; *Paul*, 2004]. Elevations above 1500 m a.s.l. were computed from the DTM from the Shuttle Radar Topography Mission (described in *Rabus et al.* [2003] and available at ftp://e0mss21u.ecs.nasa.gov/srtm/). Country names are abbreviated as follows: Austria (AT), France, (FR), Germany (DE), Italy (IT) and Switzerland (CH). The orange rectangles show the locations of the inset map (A), Figure AM6 (A) and Figure AM1 (B). The inset map shows a synthetic oblique perspective of the Aletsch region, Switzerland, generated from a DTM (*DEM25, reproduced by permission of Swisstopo, BA057338*) overlaid with a fusion of satellite images from Landsat TM (1998) and IRS-1C (1997). The Great Aletsch Glacier retreated about 2550 m from 1850 (red lines) to 1973 (blue lines) and another 680 m by 2000.

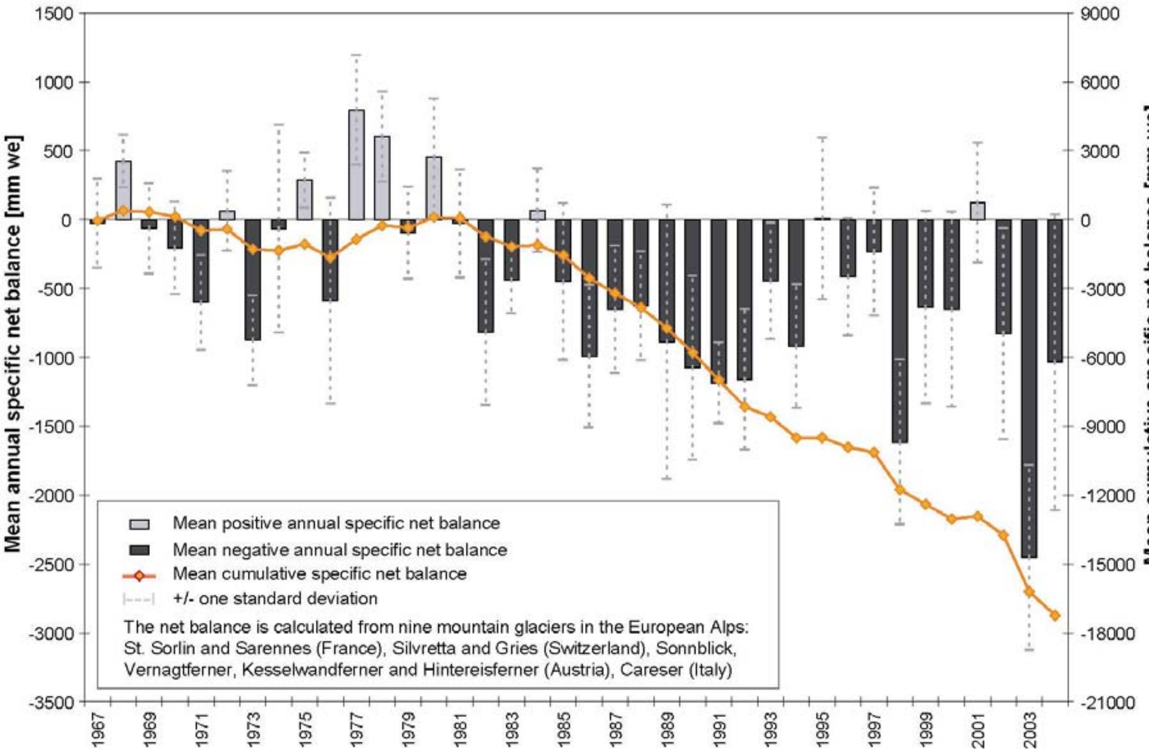

Figure AM3. Mean annual/cumulative specific net balance from nine Alpine glaciers with continuous measurements since 1967. The ice loss between 1967 (1981) and 2004 cumulated to almost 18 m w.e. The corresponding mean annual ice loss over this period(s) amounts to 0.45 (0.72) m w.e. with a mean standard deviation of 0.49 (0.55) m w.e. Maximum annual ice losses reached 2.5 m w.e. in 2003 and 1.6 m w.e. in 1998. Data source: World Glacier Monitoring Service (www.wgms.ch).

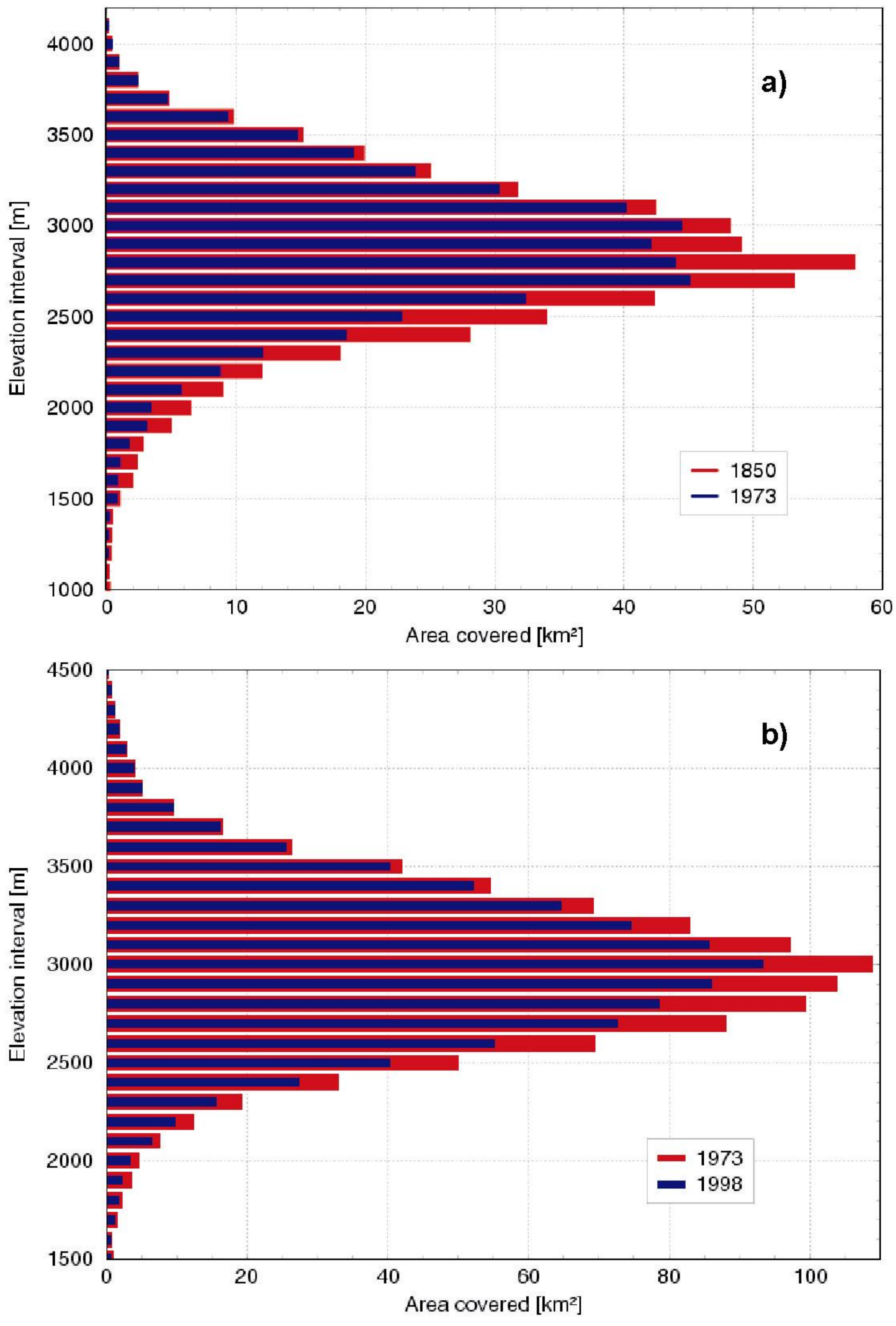

Figure AM4. Area-elevation distributions as obtained for two sub samples and time periods. The period of 1850–1973 (a) encloses 506 glaciers, while the period of 1973–1998 (b) is based on a larger sample of 713 glaciers. In order to avoid an overestimation of glacier covered areas at low elevations, we used a reconstructed DTM from 1850 for the 1850 outlines, and two DTM with 25 m resolution (*DEM25, reproduced by permission of Swisstopo, BA057338*): level1 (approx. 1985) for 1973, and level2 (approx. 1995) for 1998 outlines. All samples include the Aletsch region, which covers the full spectrum of Alpine glacier sizes, aspects, slopes and elevation ranges.

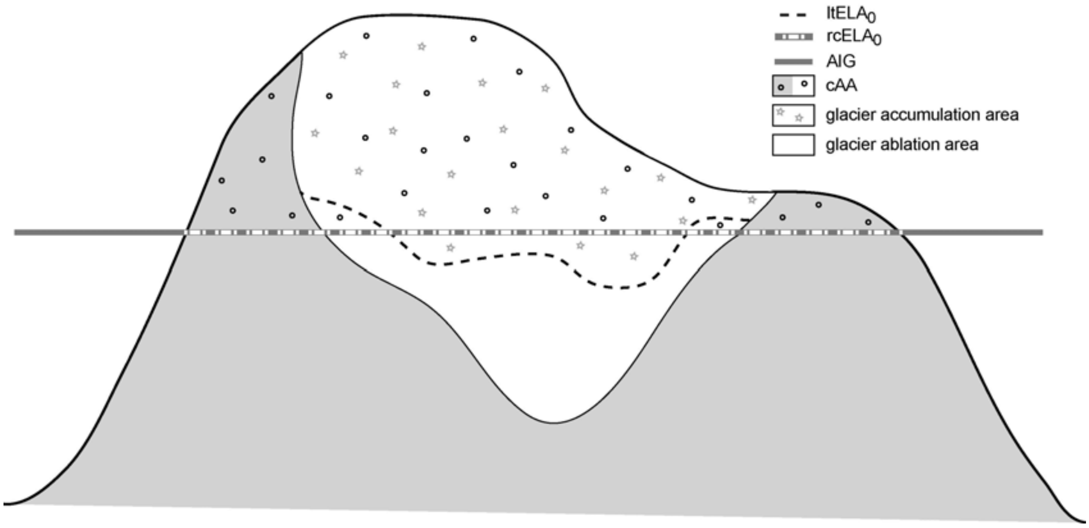

Figure AM5. Diagram to explain the concept of the altitude of instantaneous glacierization (AIG) and its association with the local topographic ELA_0 ($ltELA_0$), the regional climatic ELA_0 ($rcELA_0$) and the climatic accumulation area (cAA). Figure from [*Zemp et al.*, accepted], modified after [*Lie et al.*, 2003].

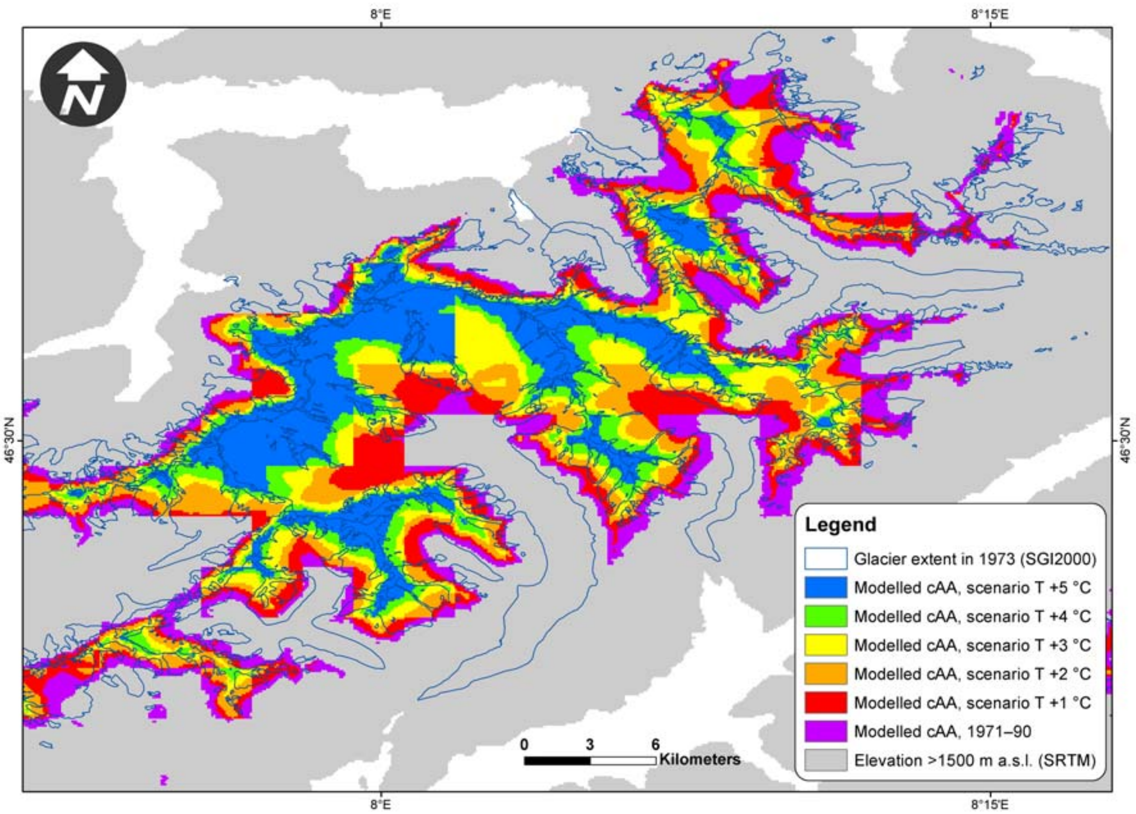

Figure AM6. Modeled climatic accumulation area (cAA) of the reference period (1971–1990) and of scenarios runs, assuming a rise in 6-month summer temperature of +1 to +5 °C. The background map shows the Aletsch region with the 1973 outlines from the Swiss Glacier Inventory [*Kääb et al.*, 2002; *Paul et al.*, 2002; *Paul*, 2004]. Elevations above 1500 m a.s.l. are computed from the SRTM3 (described in *Rabus et al.* [2003] and available at ftp://e0mss21u.ecs.nasa.gov/srtm/). For geographical location see Figure AM2, rectangle A.

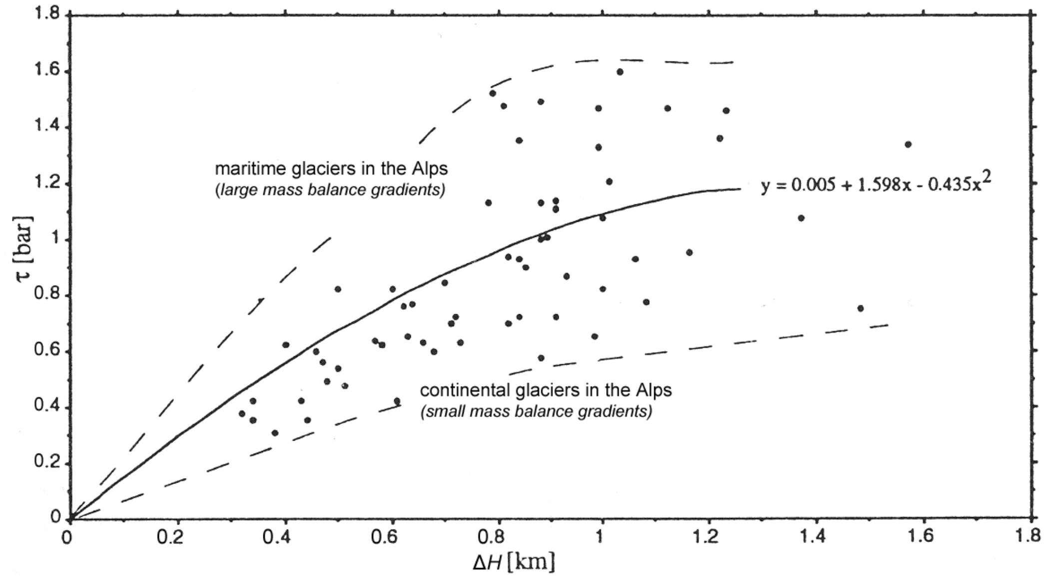

Figure AM7. Average basal shear stress along the central flowline vs altitude extent of (reconstructed late-Pleistocene) Alpine glaciers (modified after [*Haeberli and Hoelzle*, 1995].

Table AM1. Alpine glacierization in 1850, the 1970s and in 2000. §Alpine glacier area in 1850 and 2000 is extrapolated from the glacier area in the 1970s [*Haeberli et al.*, 1989] and relative area changes of the seven glacier size classes in Switzerland [*Kääb et al.*, 2002; *Paul et al.*, 2002; *Paul*, 2004]. *The relative area changes in Switzerland are calculated from the comparable sub samples, i.e., 1567 glaciers for 1850–1973 and 938 glaciers for 1973–2000. Table from *Zemp et al.* [in press b].

	Switzerland (SGI2000)								Alps			
Class	**1850**		**1973**		**2000**		**1850–1973***	**1973–2000***	**1970's**		**1850§**	**2000§**
[km²]	num.	area [km²]	num.	area [km²]	num.	area [km²]	area change [%]	area change [%]	num.	area [km²]	area [km²]	area [km²]
< 0.1	297	17.3	1022	40.1	164	3.6	-55.4	-64.6	1953	100.7	225.5	35.6
0.1–.5	715	181.3	673	153.9	448	60.3	-52.9	-45.6	2254	497.0	1055.0	270.4
0.5–1	249	172.5	151	104.1	131	63.5	-44.3	-29.1	430	299.8	538.0	212.6
1–5	253	524.4	157	296.0	141	217.1	-33.2	-17.9	425	862.3	1291.1	707.9
5–10	26	195.5	35	249.4	36	232.6	-19.7	-10.8	66	461.7	574.8	412.1
10–20	18	259.9	14	216.3	13	192.8	-14.8	-8.2	27	387.9	455.1	356.1
> 20	9	270.5	5	225.9	5	213.0	-12.3	-5.7	7	293.6	334.8	276.9
Total	1567	1621.4	2057	1285.7	938	982.9	-27.1	-16.1	5162	2902.9	4474.3	2271.6

Remarks regarding volume estimations

Changes in glacier volume were calculated by multiplying representative mass balance values with the average surface area (the mean areas at the beginning and end) of a given time period. Thereto, average mass balance values as reconstructed from cumulative glacier length changes (-0.25 m w.e. per year; [*Haeberli and Hoelzle*, 1995]) and from mass balance measurements (-0.5 m w.e. per year; provided by World Glacier Monitoring Service) were used for the periods 1850–1975 and 1975–2000, respectively. Total Alpine glacier volumes are more difficult to obtain, because characteristic mean thickness must be estimated for numerous glaciers with highly variable geometry. Straightforward volume-area scaling is simple but problematic, because it relates a variable (area) with itself (area within volume) and, hence, masks the large scatter (about ±30%) in glacier thickness as related to glacier area [*Müller et al.*, 1976].

This scatter is due to the still limited number of accurate and complete geophysical determinations of values for mean glacier thickness but also to the fact that area is not really the factor physically influencing mean glacier thickness. Mean glacier thickness as averaged over the corresponding surface area can even increase with decreasing glacier area where glaciers lose their thinnest parts (for instance, glacier tongues on steep slopes). A more process-oriented approach ([*Haeberli and Hoelzle*, 1995], based on [*Nye*, 1952] and [*Paterson*, 1981]) computes mean glacier thickness (h) along the central flowline from mean surface slope (α) as related to the average shear stress (τ):

$$h = \tau / [f \cdot \delta \cdot g \cdot \sin(a)]$$

where f is a shape factor, δ is density and g is acceleration due to gravity. It uses the fact that overall driving stress (averaged over the total length of a glacier) depends on the mass turnover and mass flux within a glacier as defined by the product of the mass balance gradient times the altitudinal extent of the glacier (Figure AM7).

According to the power-type flow law of ice [*Paterson*, 1981], mean driving stress increases from zero to a rather constant upper-limit value of about 150 kPa (1.5 bar) at altitudinal ranges (ΔH) of 500 metres and more in relatively humid climates with high mass balance gradients ([*Haeberli and Hoelzle*, 1995]; for the relation between the continentality of the climate and the mass balance gradient, see [*Kuhn*, 1981] and [*Oerlemans*, 2001]). While in the present study volume changes were estimated from average mass balances values and average surface area of a given time period, relative volume changes as related to relative changes in mean glacier thickness and area could also be approximated from:

$$\Delta V = V_2 / V_1 = \Delta A \cdot (a_1 / a_2) \cdot C$$

where slopes (a) can be calculated from glacier length and altitude range and C is a correction factor, taking into account the changing overall driving stress (τ) in the changing proportion of small glaciers (with ΔH < 500 m) in the total sample (roughly 20% for present-day conditions).

Part C: Scripts

AAR.aml

```
/* ==========================================================================
/*
/* AAR.AML
/*
/* Usage: &RUN AAR.aml <glacier> <id_field> <dem> <aar>
/*
/* Author: Michael Zemp
/*
/* Date: 23.11.2005
/* Last changes: 07.12.2005
/*
/* Description: AAR.aml computes ELA-values from glacier outlines,
/* a DEM and a defined AAR-value
/*
/* Scripted for ArcGIS Workstation 9.1 using Arc Macro Language (AML)
/*
/* Input:
/* <glacier>  glacier outlines [shapefile]
/* <id_field> fiel containing glacier ID
/* <dem>      digital elevation model [integer GRID]
/* <aar>      accumulation area ratio [%,e.g. 77%]
/*
/* Output:
/* aar<aar>_log_file.txt  log file with input variables and run time info
/* aar<aar>_ela_table.txt output table with glacier id and ELA values
/*
/* Remarks:
/* <dem> must be an integer GRID, because calculation is based on .vat
/*
/* ==========================================================================

/* --------------------------------------------------------------------------
/* Variable declarations and presettings
/* --------------------------------------------------------------------------

&args para$glacier para$id_field para$dem para$aar
&severity &error &routine bailout
PRECISION double double

/* Benutzereingaben testen...
&if [null %para$glacier%] &then
       &return &error Input parameter <glacier> is missing! ~
    USAGE: &RUN AAR.aml <glacier> <id_field> <dem> <aar>
&if [null %para$id_field%] &then
       &return &error Input parameter <id_field> is missing! ~
    USAGE: &RUN AAR.aml <glacier> <id_field> <dem> <aar>
&if [null %para$dem%] &then
       &return &error Input parameter <dem> is missing! ~
    USAGE: &RUN AAR.aml <glacier> <id_field> <dem> <aar>
&if [null %para$aar%] &then
       &return &error Input parameter <aar> is missing! ~
    USAGE: &RUN AAR.aml <glacier> <id_field> <dem> <aar>

/* Test if VAT of DEM exists
&if [exists %para$dem%.vat -info] &then
       &type %para$dem%.vat exists!
&else
       &return &error ERROR: %para$dem%.vat does not exist!
```

```
/* --------------------------------------------------------------------------
/* create log file
/* --------------------------------------------------------------------------

&type Create log file...

/* Set start date and time
&sv para$startdate = [DATE -CAL]
&sv para$starttime = [DATE -TIME]

/* create log file and open in write mode
&sv para$unit = [OPEN aar%para$aar%_log_file.txt para$openstat -WRITE]
&if %para$openstat% eq 0 &then
 &type aar%para$aar%_log_file.txt opened successfully!
&else &type Error: openstat = %para$openstat% (ESRI 1993: p. 10-25)

/* write header to file
&sv para$writestat = [WRITE %para$unit% [QUOTE -------------------------]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Log File for AAR.aml]]
&sv para$writestat = [WRITE %para$unit% [QUOTE -------------------------]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Start Date: %para$startdate%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Start Time: %para$starttime%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE ---]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Input Glacier Shp:
%para$glacier%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Input ID_Field:
%para$id_field%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Input DEM: %para$dem%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Input AAR: %para$aar%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE Output Table:
aar%para$aar%_ela_table.txt]]
&sv para$writestat = [WRITE %para$unit% [QUOTE ---]]

&if %para$writestat% eq 0 &then
 &type Header written successfully!
&else &type Error: writestat = %para$writestat% (ESRI 1993: p. 10-25)

/* close text file
&sv para$closestat = [CLOSE %para$unit%]
&if %para$closestat% eq 0 &then
 &type File closed successfully!
&else &type Error: closestat = %para$closestat% (ESRI 1993: p. 10-25)


/* --------------------------------------------------------------------------
/* vector-2-raster conversion
/* --------------------------------------------------------------------------

&type vector-2-raster conversion...

/* start GRID
GRID

/* Set cellsize to resolution of <dem>
SETCELL %para$dem%
&type Cell size for this analysis is set to: [SHOW SETCELL]

/* convert shapefile <glacier> to a raster
glaciergrid = SHAPEGRID(%para$glacier%, %para$id_field%)

&type v2r done!
```

```
/* ---------------------------------------------------------------------------
/* create output table
/* ---------------------------------------------------------------------------

&type Create output table...

/* create text file and open in write mode
&sv para$unit = [OPEN aar%para$aar%_ela_table.txt para$openstat -WRITE]
&if %para$openstat% eq 0 &then
 &type aar%para$aar%_ela_table.txt opened successfully!
&else &type Error: openstat = %para$openstat% (ESRI 1993: p. 10-25)

/* write header to file
&sv para$writestat = [WRITE %para$unit% [QUOTE Glacier_ID ELA]]
&if %para$writestat% eq 0 &then
 &type Header written successfully!
&else &type Error: writestat = %para$writestat% (ESRI 1993: p. 10-25)

/* close text file
&sv para$closestat = [CLOSE %para$unit%]
&if %para$closestat% eq 0 &then
 &type File closed successfully!
&else &type Error: closestat = %para$closestat% (ESRI 1993: p. 10-25)


/* ---------------------------------------------------------------------------
/* Create list of all glacier IDs
/* ---------------------------------------------------------------------------

&type Create list of all glacier IDs

/* define valiable to hold all glacier-IDs
&sv para$list
&type ID-list: %para$list%

CURSOR cur DECLARE glaciergrid.vat INFO
CURSOR cur OPEN

&messages &off
&do &while %:cur.AML$NEXT%
 &sv para$list = %para$list% %:cur.value%
 CURSOR cur NEXT
&end /* of &do&while
&messages &on

CURSOR cur REMOVE

&type glacier-ID-list: %para$list%

&type glacier-ID-list created!


/* -*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*
/* Iteration for each glacier
/* -*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*

&type --------------------------------------------------------
&type Iteration for each glacier...

&do current_glacier &list %para$list%
 &type """""""""""""""""""""""""""""""""""""""""""""""
 &type loop for glacier no. %current_glacier%
```

```
/* --------------------------------------------------------------------------
/* Calculate ELA of current glacier
/* --------------------------------------------------------------------------

&type Calculate ELA of current glacier

/* create mask of current glacier
currentglac = SELECT(glaciergrid, [quote value eq %current_glacier%])

/* get DEM values for current glacier
glacdem_cur = int(SELECTMASK(%para$dem%, currentglac))

/* calculate area of current glacier (in no. of grid cells)
CURSOR cur DECLARE currentglac.vat INFO
CURSOR cur OPEN
&sv para$glacier_area = %:cur.count%
&type Glacier area: %para$glacier_area%
CURSOR cur REMOVE

/* calculate area of ablation area (in no. of grid cells)
&sv para$abl_area [CALC %para$glacier_area% * ~
[CALC [CALC 100 - %para$aar%] / 100]]
&type Size of ablation area: %para$abl_area%

/* Calculate ELA
&sv para$h_ela = 0
&sv para$cellcount = 0
&sv condition = .true.

CURSOR cur DECLARE glacdem_cur.vat INFO
CURSOR cur OPEN

&messages &off
&do &while %condition%
 &sv para$h_ela = %:cur.value%
 &sv para$cellcount = %para$cellcount% + %:cur.count%
 CURSOR cur NEXT
 &if %para$cellcount% ge %para$abl_area% &then
      &sv condition = .false.
&end /* of &do&while
&messages &on

CURSOR cur REMOVE

/* --------------------------------------------------------------------------
/* Write ID and ELA of current glacier to resulting table
/* --------------------------------------------------------------------------

&type Write ID and ELA of current glacier to resulting table

&type ELA of glacier no. %current_glacier%: %para$h_ela%

/* --------------------------------------------------------------------------
/* Add results to output table
/* --------------------------------------------------------------------------

&type Add current Glacier_ID and ELA to output table...

/* open text file in append mode
&sv para$unit = [OPEN aar%para$aar%_ela_table.txt para$openstat -APPEND]
&if %para$openstat% eq 0 &then
 &type aar%para$aar%_ela_table.txt opened successfully!
&else &type Error: openstat = %para$openstat% (ESRI 1993: p. 10-25)

/* write line with results to file
&sv para$writestat = [WRITE %para$unit% [QUOTE %current_glacier%
%para$h_ela%]]
&if %para$writestat% eq 0 &then
 &type Line appended successfully!
&else &type Error: writestat = %para$writestat% (ESRI 1993: p. 10-25)
```

```
 /* close text file
 &sv para$closestat = [CLOSE %para$unit%]
 &if %para$closestat% eq 0 &then
  &type File closed successfully!
 &else &type Error: closestat = %para$closestat% (ESRI 1993: p. 10-25)

 /* delete temporary datasets
 kill currentglac
 kill glacdem_cur

 &type End of loop for glacier no. %current_glacier%
 &type """""""""""""""""""""""""""""""""""""""""""""""""

&end /* of &do&list

&type End of iteration!
&type -------------------------------------------------------

/* -*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*
/* End of iteration
/* -*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*-*

/* Quit GRID
QUIT


/* ----------------------------------------------------------------------------
/* Add end time to log file
/* ----------------------------------------------------------------------------

&type Add end time to log file...

/* Set end date and time
&sv para$enddate = [DATE -CAL]
&sv para$endtime = [DATE -TIME]

/* Open log file in append mode
&sv para$unit = [OPEN aar%para$aar%_log_file.txt para$openstat -APPEND]
&if %para$openstat% eq 0 &then
 &type aar%para$aar%_log_file.txt opened successfully!
&else &type Error: openstat = %para$openstat% (ESRI 1993: p. 10-25)

/* write header to file
&sv para$writestat = [WRITE %para$unit% [QUOTE End Date: %para$enddate%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE End Time: %para$endtime%]]
&sv para$writestat = [WRITE %para$unit% [QUOTE ---]]

&if %para$writestat% eq 0 &then
 &type Header written successfully!
&else &type Error: writestat = %para$writestat% (ESRI 1993: p. 10-25)

/* close text file
&sv para$closestat = [CLOSE %para$unit%]
&if %para$closestat% eq 0 &then
 &type File closed successfully!
&else &type Error: closestat = %para$closestat% (ESRI 1993: p. 10-25)


&type AAR.aml successfully terminated!
&type -------------------------------------------------------

&return
```

```
/* ----------------------------------------------------------------------------
   &routine  BAILOUT
/* ----------------------------------------------------------------------------

&severity &error &ignore
&type BAILOUT Routine...

/* if still running, quit GRID
&if [show program] eq GRID &then QUIT

/* Variablen löschen
&dv para$*
&type 'para$'-Variables deleted!

&return &warning Error in AAR.aml!
```

MADTEN.py

```
# ############################################################################
# MADTEN.py
#
# Usage: MADTEN.py <InRaster> <OutRaster>
#
# Author: M. Zemp
#
# Date: 01.06.2005
# Last changes: 15.06.2005
#
# Description: MADTEN is a focal function calculating the mean ablolute
# difference of each cell value to the ones of its eight immediate
# neighbours in a 3x3 neighbourhood.
#
# Scripted for ArcGIS 9.1 using Pythonwin 2.1
#
# Input:
# <InRaster>  input raster on which MADTEN is to be calculated [GRID]
# <OutRaster> name of the output raster to be written [GRID]
#
# Output:
# <OutRaster> GRID of calculated MADTEN-values
#
# Remarks:
# The cell size of <InRaster> will be set as output cell size
# SpatialAnalyst-Licence is needed!
#
# ############################################################################

print "MADTEN.py"
# Import system modules
import sys, os, win32com.client

# Create the Geoprocessor object
gp = win32com.client.Dispatch("esriGeoprocessing.GpDispatch.1")

# Set the InRaster
InRaster = sys.argv[1]
print "InRaster: " + InRaster

# Set the OutRaster
OutRaster = sys.argv[2]
print "OutRaster: " + OutRaster

# Set the CellSize, from InRaster [Cheers to Michael Höck :-)]
dscInRaster = gp.Describe(InRaster)
CellSize = dscInRaster.MeanCellWidth
print "CellSize: " + str(CellSize)

# Set the Workspace for temporary data, from InRaster
Workspace = str(os.path.split(InRaster)[0]) + "\\"

print "Workspace: " + Workspace

# Allow overwriting of output
gp.OverwriteOutput = 1
print "Output overwriting is ON!"
```

```
# Check out SpatialAnalyst extension license
try:
    gp.CheckOutExtension("Spatial")
except:
    print gp.GetMessages(2)

# -----------------------------------------------------------------------------
# SHIFT each of the 8 neighbours of InRaster into the centre of the 3x3-window
# -----------------------------------------------------------------------------

try:
    print "Entering Shift-Section..."
    # Shift LowerLeft into the centre
    LL = Workspace + "LL"
    gp.Shift_management(InRaster, LL, CellSize, CellSize, InRaster)
    print "InRaster shifted and saved as LL!"

    # Shift LowerCentre into the centre
    LC = Workspace + "LC"
    gp.Shift_management(InRaster, LC, 0, CellSize, InRaster)
    print "InRaster shifted and saved as LC!"

    # Shift LowerRight into the centre
    LR = Workspace + "LR"
    gp.Shift_management(InRaster, LR, -CellSize, CellSize, InRaster)
    print "InRaster shifted and saved as LR!"

    # Shift MidLeft into the centre
    ML = Workspace + "ML"
    gp.Shift_management(InRaster, ML, CellSize, 0, InRaster)
    print "InRaster shifted and saved as ML!"

    # Shift MidRight into the centre
    MR = Workspace + "MR"
    gp.Shift_management(InRaster, MR, -CellSize, 0, InRaster)
    print "InRaster shifted and saved as MR!"

    # Shift UpperLeft into the centre
    UL = Workspace + "UL"
    gp.Shift_management(InRaster, UL, CellSize, -CellSize, InRaster)
    print "InRaster shifted and saved as UL!"

    # Shift UpperCentre into the centre
    UC = Workspace + "UC"
    gp.Shift_management(InRaster, UC, 0, -CellSize, InRaster)
    print "InRaster shifted and saved as UC!"

    # Shift UpperRight into the centre
    UR = Workspace + "UR"
    gp.Shift_management(InRaster, UR, -CellSize, -CellSize, InRaster)
    print "InRaster shifted and saved as UR!"

except:
    print "Error in Shift-Section of MADTEN.py!"
    print gp.GetMessages(2)

print "Shift-Section was successful!"
```

```
# ---------------------------------------------------------------------------
# MADTEN: Calculate mean absolute difference to eight neigbours
# ---------------------------------------------------------------------------

try:
    print "Entering MADTEN-Section..."

    # Set MapAlgebra-Expression
    InExpression = OutRaster + " = ((abs(" + InRaster + " - " + LL + ")" +\
    " + abs(" + InRaster + " - " + LC + ")" +\
    " + abs(" + InRaster + " - " + LR + ")" +\
    " + abs(" + InRaster + " - " + ML + ")" +\
    " + abs(" + InRaster + " - " + MR + ")" +\
    " + abs(" + InRaster + " - " + UL + ")" +\
    " + abs(" + InRaster + " - " + UC + ")" +\
    " + abs(" + InRaster + " - " + UR + ")" + ") / 8)"
    print "MultiOutputMapAlgebra Expression: '" + InExpression + "'"

    # Process MapAlgebra
    gp.MultiOutputMapAlgebra_sa(InExpression)

except:
    print "Error in MADTEN-Section of MADTEN.py!"
    print gp.GetMessages(2)

print "MADTEN-Section was successful!"

# ---------------------------------------------------------------------------
# CLEANUP: delete temporary files
# ---------------------------------------------------------------------------

try:
    print "Entering CLEANUP-Section..."

    # delete shifted rasters
    gp.delete(LL)
    gp.delete(LC)
    gp.delete(LR)
    gp.delete(ML)
    gp.delete(MR)
    gp.delete(UL)
    gp.delete(UC)
    gp.delete(UR)

except:
    print "Error in CLEANUP-Section of MADTEN.py!"
    print gp.GetMessages(2)

print "CLEANUP-Section was successful!"

# Well done...
print "MADTEN.py was successful!"
print ""
```

ELA.py

```
# ########################################################################
# ELA.py
#
# Usage: ELA.py <InDEM> <InTemperature> <InPrecipitation> <InPgradient>
#               <InTgradient> <InA> <InB> <RunDescription>
#
# Author: Michael Zemp
#
# Date: 07.09.2005
# Last changes: 10.06.2006
#
# Description: Computes the Altitude of Instantaneous Glacierisation (AIG),
# the climatic Accumulation Area (cAA) and the regional climatic ELA0
#
# Scripted for ArcGIS 9.1 using PythonWin 2.1
#
# Input:
# <InDEM>           digital elevation model [GRID; m]
# <InTemperature>   temperature [GRID; °C]
# <InPrecipitation> precipitation [GRID; mm]
# <InPgradient>     precipitation-elevation gradient
#                   [floating GRID or value; % per 100 m]
# <InTgradient>     temperature lapse rate
#                   [floating GRID or value; °C per 100 m]
# <InA>             constant A from emp. relation: P=A*e^(B*T)
# <InB>             constant B from emp. relation: P=A*e^(B*T)
# <RunDescription>  prefix, added to all output files and name of the
#                   logfile describing the parameters of the model run
#
# Output:
# <RunDescr>_log.txt log file with input variables and run time info
# <RunDescr>_h       altitude of AIG above terrain [GRID]
# <RunDescr>_AIG     AIG [GRID]
# <RunDescr>_cAA     cAA [GRID]
# <RunDescr>_ELA0    ELA0 [GRID]
#
# Remarks:
# <InTgradient> must be indicated as positive values,
#               e.g. 0.65 °C per 100 m
# <InPgradient> must be indicated in % per 100 m,
#               e.g. 0.08% per 100 m
# <RunDescription> must be no longer than 8 characters (GRID convention)
#
# ########################################################################

# ------------------------------------------------------------------------
# Presettings I
# ------------------------------------------------------------------------
print "ELA.py"
# Import system modules
import sys, string, os, win32com.client
from time import gmtime, strftime

# Print starting time
DateTime = strftime("%a, %d %b %Y %H:%M:%S")
print DateTime

# Create the Geoprocessor object
gp = win32com.client.Dispatch('esriGeoprocessing.GpDispatch.1')
```

```
# ---------------------------------------------------------------------------
# User input
# ---------------------------------------------------------------------------
# Set input variables
InDEM = sys.argv[1]
print "DEM: " + InDEM

InTemperature = sys.argv[2]
print "Temperature: " + InTemperature

InPrecipitation = sys.argv[3]
print "Precipitation: " + InPrecipitation

InPgradient = sys.argv[4]
print "Precipitation gradient: " + InPgradient

InTgradient = sys.argv[5]
print "Temperature gradient: " + InTgradient

InA = sys.argv[6]
print "A: " + InA

InB = sys.argv[7]
print "B: " + InB

RunDescr = sys.argv[8]
print "Label of current model run: " + RunDescr

# ---------------------------------------------------------------------------
# Presettings II
# ---------------------------------------------------------------------------

# Set the Workspace for temporary data, from InTemperature
Workspace = str(os.path.split(InTemperature)[0]) + "\\"
gp.Workspace = Workspace
print "Workspace is set to: " + Workspace

# Allow overwritting of output
gp.OverwriteOutput = 1
print "Output overwriting is ON!"

# Check out SpatialAnalyst extension license
try:
    gp.CheckOutExtension("Spatial")

    # Set the CellSize, from InDEM [Cheers to Michael Höck :-)]
    dscInDEM = gp.Describe(InDEM)
    CellSize = dscInDEM.MeanCellWidth
    gp.CellSize = CellSize
    print "CellSize is set to: " + str(CellSize)

except:
    print gp.GetMessages(2)
```

```
# ---------------------------------------------------------------------------
# LOG-File
# ---------------------------------------------------------------------------
print "Create LOG-file..."

log_file = Workspace + RunDescr + "_log.txt"
f = open(log_file, "w")
print f

# Write UserInput to LOG-File
f.write("Start of model run: " + DateTime + "\n")
f.write("\n")
f.write("InDEM: " + InDEM + "\n")
f.write("InTemperature: " + InTemperature + "\n")
f.write("InPrecipitation: " + InPrecipitation + "\n")
f.write("InPgradient: " + InPgradient + "\n")
f.write("InTgradient: " + InTgradient + "\n")
f.write("InA: " + InA + "\n")
f.write("InB: " + InB + "\n")
f.write("\n")
f.close()
print f
print "LOG-File created and input values listed!"

# ---------------------------------------------------------------------------
# Calculation of h
# ---------------------------------------------------------------------------
print "Calculating h..."
# Set InExpression for MOMA calculating h
InExpression = Workspace + RunDescr + "_h = (ln(" + InA +\
") + " + InB + " * " + InTemperature + " - " + "ln(" +\
InPrecipitation + ")) / (((ln(100 + " + InPgradient +\
")) / 100) + (" + InB + " * " + InTgradient + ")) * 100"
print "MultiOutputMapAlgebra Expression for h: " + InExpression

# Calculate h
try:
    # Process: MapAlgebra
    print "Start MOMA..."
    gp.MultiOutputMapAlgebra_sa(InExpression)
    print "MOMA done!"

except:
# Print error message if an error occurs
    gp.GetMessages()

print "Calculation of h terminated!"

# ---------------------------------------------------------------------------
# Calculation of AIG
# requires InDEM, h
# ---------------------------------------------------------------------------
print "Calculating AIG..."
# Set InExpression for MOMA calculating AIG
InExpression = Workspace + RunDescr + "_AIG = " + InDEM +\
" + " + Workspace + RunDescr + "_h"
print "MultiOutputMapAlgebra Expression for AIG: " + InExpression

# Calculate AIG
try:
    # Process: MapAlgebra
    print "Start MOMA..."
    gp.MultiOutputMapAlgebra_sa(InExpression)
    print "MOMA done!"

except:
# Print error message if an error occurs
    gp.GetMessages()

print "Calculation of AIG terminated!"
```

```
# ----------------------------------------------------------------------------
# Calculation of cAA
# requires h
# ----------------------------------------------------------------------------
print "Calculating cAA..."
# Set InExpression for MOMA calculating cAA
InExpression = Workspace + RunDescr + "_cAA = con(" +\
Workspace + RunDescr + "_h <= 0, 1, 0)"
print "MultiOutputMapAlgebra Expression for cAA: " + InExpression

# Calculate cAA
try:
    # Process: MapAlgebra
    print "Start MOMA..."
    gp.MultiOutputMapAlgebra_sa(InExpression)
    print "MOMA done!"

except:
# Print error message if an error occurs
    gp.GetMessages()

print "Calculation of cAA terminated!"

# ----------------------------------------------------------------------------
# Calculation of ELA0
# requires cAA
# ----------------------------------------------------------------------------
print "Calculating ELA0..."
# Procedure:
# - expanding the cAA by 1 cell
# - subtracts cAA
# - sets all cells to NoData that are equal to 0 (i.e. the expanded cAA
# boundary gets the value 1 = ELA0)

# Set InExpression for MOMA calculating ELA0
InExpression = Workspace + RunDescr + "_ELA0 = SETNULL(((EXPAND(" +\
Workspace + RunDescr + "_cAA, 1, LIST, 1)) - " + Workspace + RunDescr +\
"_cAA) EQ 0, 1)"
print "MultiOutputMapAlgebra Expression for ELA0: " + InExpression

# Calculate ELA0
try:
    # Process: MapAlgebra
    print "Start MOMA..."
    gp.MultiOutputMapAlgebra_sa(InExpression)
    print "MOMA done!"

except:
# Print error message if an error occurs
    gp.GetMessages()

print "Calculation of ELA0 terminated!"
```

```
# ---------------------------------------------------------------------------
# Finish
# ---------------------------------------------------------------------------
# Print ending time
DateTime = strftime("%a, %d %b %Y %H:%M:%S")
print DateTime

# Add end-time to LOG-File

f = open(log_file, "a")
print f

# Write UserInput to LOG-File
f.write("End of model run: " + DateTime)
f.close()
print f

# End of script
print "Successfully reached end of script :-)"
print "--------------------------------------"
```